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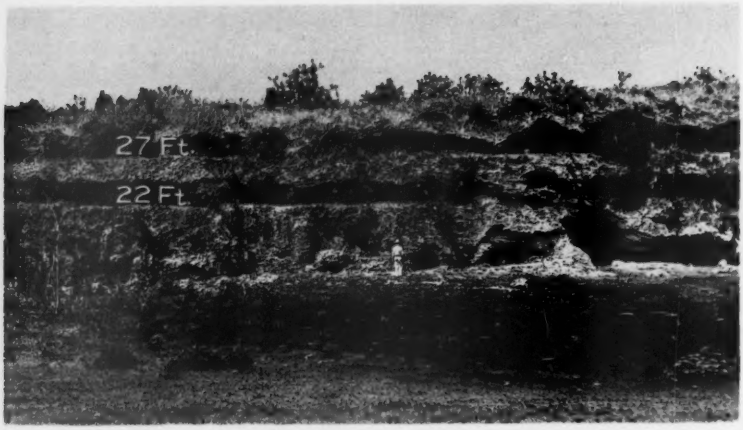

FIGURE 1. PARALLEL NOTCHES 22 AND 27 FEET ABOVE SEA LEVEL AT KAILUA, OAHU
Cut in lithified dunes during the Waimanalo stand of the sea. Note the sea caves.

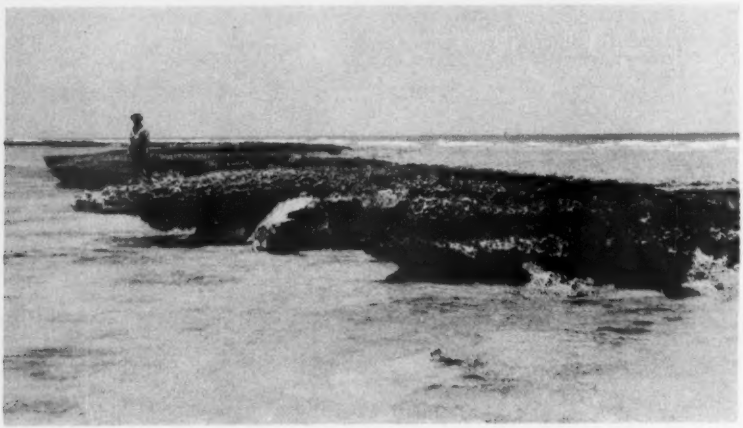

FIGURE 2. EMERGED REEF OF THE 5-FOOT SEA, MIDWAY ATOLL
Wide platform on the right is chiefly a result of the sea beveling the emerged reef to present sea level. Note the notch and visor under the man on the lagoonal side of the reef. No fresh water table exists.

EMERGED SHORE NOTCHES ON OAHU AND MIDWAY ISLANDS, HAWAII

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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SHORE BENCHES ON NORTH PACIFIC ISLANDS

BY HAROLD T. STEARNS

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ABSTRACT

Confusion exists in the literature about the three low benches on Pacific Islands, because height alone is not an index of the ocean level which made them. The lowest bench is now being made by erosion and weathering. The other two were made when the sea stood about 5 feet and 25 feet higher. These higher benches where present have usually been reduced by weathering and erosion and that of the 5-foot sea is readily confused with the present sea bench. The lowest bench, where completely developed, has five components: (1) channeled ramp, (2) *lithothamnium* ridge, (3) platform, (4) notch, and (5) beach deposits. Benches are described on Guam, Midway, and Oahu Islands and their concordance in height suggests a eustatic origin. A detailed survey of the benches at Amantes Point, Guam, is given.

THE PROBLEM

The three low benches bordering the shores of Pacific Islands have been variously interpreted throughout the existing literature. The lowest one has been described as the shore platform (Bartrum, 1926) and the solution bench (Wentworth, 1939). The upper two have been variously described as the eustatic bench (Wentworth and Palmer, 1925), the

2-meter bench (Johnson, 1931), the water-level bench (Wentworth, 1938), the bench of a 12-foot stand of the sea (Wentworth, 1926, p. 120), the bench of a 6-meter stand of the sea (Daly, 1934, p. 157), and the storm-wave bench (Bartrum, 1924). The writer has spent the last 10 years on Pacific Islands, including the Hawaiian Islands, Midway, Guam, Philippines, Samoa, Fiji, New Zealand, Australia, Java, and Japan and has seen many of the specific bench localities. It is not surprising to find the reports unreconciliable since the problem is so complex that an incorrect interpretation is readily made from the study of a single locality. The result is a great lack of precise data on action of the sea in all its moods on island shores, particularly on coral strands.

Confusion arises from the fact that these three benches, made by three different stands of the sea, may exist on shores that have been stable since late Pleistocene. Height alone is a poor indicator of the stand, because the bench of any one stand may vary in height an amount equal or nearly equal to the interval between the successive halts of the sea. Reefs complicate the problem on coral islands as they are constructional features. Lowering of relic benches by weathering and storm-wave erosion also complicates the subject.

The various stands are clearly separated from each other in some places on Oahu Island, Hawaii, where the notches and benches of the several seas were made in laminated, cross-bedded, lithified coral sand dunes. The dunes have a uniform texture and grain size. Thus, the effects of bedding planes, hard and soft layers, coral-reef growth, and other complications are eliminated. Oahu was stable during late Pleistocene, as shown by several horizontal older emerged reefs (Stearns, 1935c). The clear record shows that the sea halted for a short time 27 feet above present sea level and then dropped to 22 feet (Pl. 1, fig. 1). Considerable reefs grew during these two halts. The two stands are separated with difficulty in most localities; hence they will collectively be called hereafter the 25-foot stand (Stearns, 1935c, p. 1944). Next the sea dropped to 5 feet above present sea level and halted there long enough to produce a bench from a few feet to several hundred feet in width (Fig. 1; Pl. 3, fig. 1). Little coral grew during this halt, indicating either a shorter pause or existence of colder water than at the 25-foot level. Finally, the sea receded to its present level. It has already produced a bench or shore platform comparable in width to that of the 5-foot sea. This present shore platform ranges from a few inches to about 12 feet above sea level (Stearns, 1935b).

Possibly, under fortuitous circumstances or in weak rocks, notches indicating short intermediate pauses of the sea were made. Traces of a short halt were found by the writer 8 feet above sea level in Tumon Bay,

FIGURE 1. THE SHORE PLATFORM
(1) Channeled ramp; (2) *Lithothamnium* ridge; (3) platform with patches of sand and potholes; (4) notch with visor. (Photo by S. Higgins)

FIGURE 2. CHANNELED RAMP AT OUTER EDGE OF THE SHORE PLATFORM
(Photo by S. Higgins.)

FIGURE 3. TERRACES CONSTRUCTED BY CORALLINE ALGAE MAKE THE *Lithothamnium* RIDGE ON THE SHORE PLATFORM

PRESENT SHORE PLATFORM NEAR CATALINA POINT, GUAM

FIGURE 1. WINDWARD SIDE OF AMANTES POINT
Man to left stands on the pitted 5-foot bench. Man to right stands opposite the point and on the reef flat which is covered at high tide.

FIGURE 2. LEEWARD SIDE OF AMANTES POINT
Showing abrupt drop in the shore platform as a result of decreased height of waves rounding the point.

FIGURE 3. NOTCHES AT NATON POINT, TUMON BAY, GUAM
Lower notch is being made by the present sea, the upper notch was formed by a sea 5 feet higher.

Guam. Wentworth (1926, p. 120; Wentworth and Hoffmeister, 1939, p. 1566) believes that the sea halted at the 12-foot level on its way down, but the notch described by him is too low and too poorly developed to be satisfactory evidence of this pause.

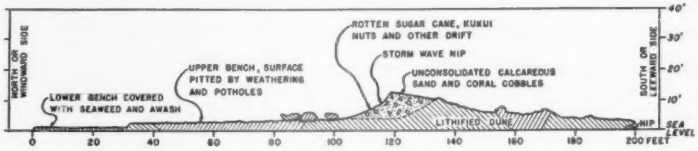

FIGURE 1.—*Cross section of Kipapa Island in Kaneohe Bay, Oahu*
Showing shore platform, nip of the present sea, and bench of the 5-foot sea.

BENCH OR SHORE PLATFORM OF THE PRESENT SEA

The origin of the several benches made by past seas may be inferred from study of the bench or shore platform being formed by the present sea. On limestone shores in tropical seas the typical fully developed platform has the following units progressing landward (Fig. 2, insert B, section 4; Pl. 2, fig. 1):

(1) A channeled ramp, lying below mean sea level and bare only during the recession of large waves or during extremely low tides (Pl. 2, fig. 2). The ramp is exposed to intense wave action and is covered with seaweed. Large blocks may be quarried from its face during storms.

(2) A *lithothamnium* ridge built of coralline algae and reaching several feet above mean sea level on exposed coasts. It decreases in height inland and is identified by delicate rimmed terraces (Pl. 2, fig. 3).

(3) A platform always awash except during low tide in calm seas. It is pitted in many places with potholes, channeled with shallow runways most of which run parallel to the shore, and overlain with scattered small patches of sand (Pl. 2, fig. 1). Where a *lithothamnium* ridge is missing many more algae are found on the bench.

(4) A notch where the most vigorous abrasion and solution is taking place. A visor is usually developed over the notch in quiet bays.

(5) Storm waves leave deposits of boulders and subrounded blocks on exposed coasts and sand on sheltered coasts 5 to 10 feet above mean sea level if the coasts are not too steep.

The channeled ramp and *lithothamnium* ridge may be poorly developed on sheltered coasts, especially the latter in high latitudes. (See Fig. 2, insert B, sections 1, 2.) The platform rises from about sea level on sheltered coasts to 12 feet on exposed headlands. Its width ranges from a few to several hundred feet.

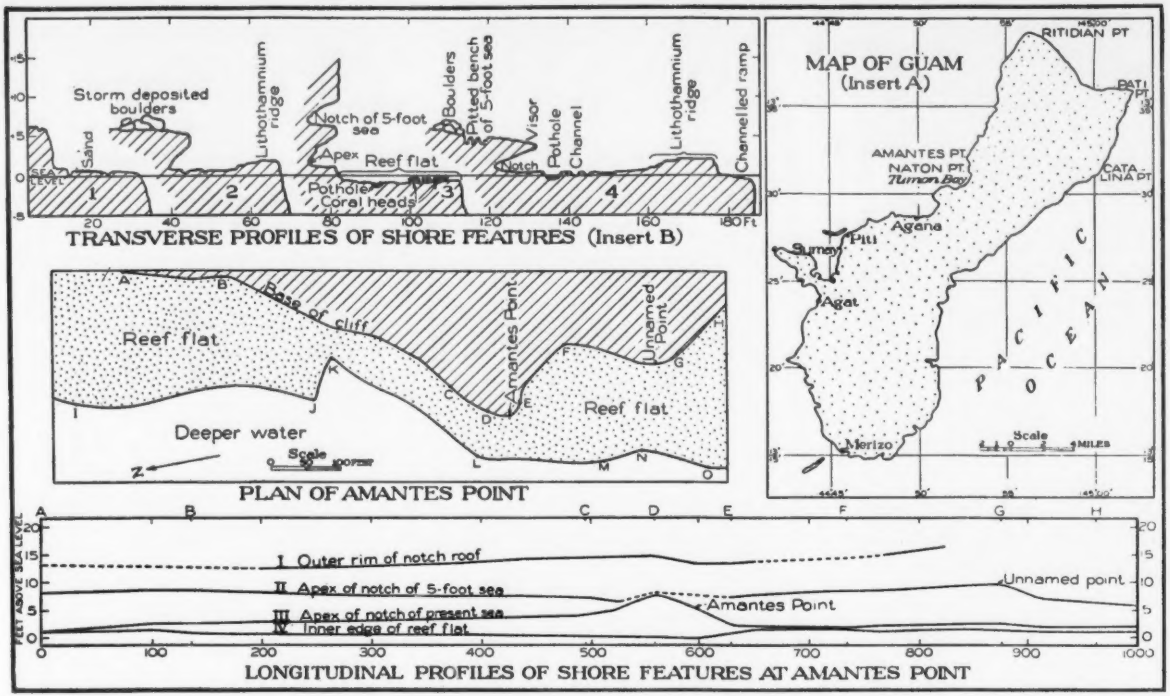

FIGURE 2.—*Plan and profiles of shore features, Amantes Point, Guam*

Broken lines in longitudinal profiles indicate parts eroded away. Profile II decreases in height between B and C, is missing between C and E, and is abnormally low at H because of subsequent erosion and weathering. Insert A shows map of Guam. Insert B shows transverse profiles of shore features on limestone coasts. (1) Simple platform; (2) platform with *lithothamnium* ridge and notch; (3) platform typified by Amantes Point with weathered bench remnant and notch of 5-foot sea, present sea bench and notch, and reef flat with potholes near shore and live coral heads at outer margin; and (4) typical bench with all characteristic features.

ORIGIN OF THE SHORE PLATFORM

The platform is made by both mechanical and chemical action. Sufficient sand is tossed upon the bench by rough seas for abrasion to be a constant factor, but solution by sea water is also an active process. The quantity of limestone that may be dissolved by sea water depends upon such factors as changes in temperature, organic acids from decomposing plant and animal life, and fresh water diluting the ocean water during heavy rains. These factors are small but ever present. Reef limestone is massive, and joints are scarce, hence the solution and abrasion processes making the platform keep ahead of the quarrying at its outer face. Recently it has been proposed that such platforms are produced by the coalescence of rain-water solution pits in an older higher bench (Wentworth, 1939, p. 21-22). There can be no doubt that solution pits in the higher bench reduce to some extent the amount of rock to be removed by the waves in producing the platform, but there are miles of coast line on Pacific isles where the higher platform is not appreciably pitted by solution and where the notch is advancing into a vertical face of limestone (Pl. 3, fig. 1). Also, the platforms rise toward unprotected headlands, a state obviously conditioned by the size of the waves reaching the shore and not by the amount of rain or the number of solution pits in the upper bench (Fig. 2).

Solution pits made by rain water cannot account for notches with visors as the pits do not penetrate the visor and consequently could not form the platform. Wentworth advanced the theory for such places that ground water escaping at the notch level dissolves the limestone (1939, p. 23). Many small pedestal rocks stand in lagoons surrounded with a notch and visor where no fresh water could possibly exist,—as on the emerged atoll at Midway (Pl. 1, fig. 2). Such notches must be made by the sea and not by ground water.

The writer agrees with Johnson (1931, p. 3) that the platform is made by the present sea and also that such platforms

> "normally have their inner margins from a few centimeters up to two meters or more above the level of ordinary high tides, the exact level varying with conditions of exposure to storm waves, tidal range, breadth of platform and other local conditions. The cutting is effective in unweathered rock, and usually is independent of the ground-water level."

Johnson also points out convincingly that such platforms have been incorrectly interpreted as having been produced 10 to 12 feet below sea level and are therefore indicative of a recent drop in sea level of about 12 to 15 feet. This misinterpretation of sea level may have arisen because abandoned sea caves and shore deposits of a higher sea have been found, and

typical bench with all characteristic features.

the platform was wrongly supposed to have been formed contemporaneously. The sea caves and shore deposits belong either to the older 5-foot or 25-foot stands of the sea. However, Johnson erred in assuming that the present shore platform has always been referred to by writers as a "2-meter bench." Several authors clearly describe the bench abandoned by the 5-foot sea. The present writer is convinced that the term 2-meter bench applies more accurately to the abandoned bench of the 5-foot sea than to the present platform. The abandoned 5-foot bench has in places been partly dressed down by water-level weathering (Wentworth, 1938) and storm-wave erosion until it lies intermediate in level between the unmodified bench of the 5-foot sea and the shore platform. At Amantes Point, Guam, the features and relations of the shore platform and of the bench of the 5-foot sea, one below the other, are excellently displayed.

BENCHES AT AMANTES POINT, GUAM

Amantes Point lies on the leeward coast of Guam and can be readily reached from Tumon Bay. A detailed transit survey of this point and profiles of the benches and notches are shown in Figure 2. The point is exposed to strong wave action during storms. A nearly vertical cliff of limestone 300 feet high, indented by a deep notch, is skirted by a living reef (Pl. 3, fig. 1). That the inner part of the reef flat is cut into the limestone forming the cliff is proved by joint cracks which may be traced from the flat into the cliff. Further, the veneer of living coral disappears near the base of the cliff, and abraded limestone is exposed. A notched bench abandoned by the 5-foot stand of the sea and now deeply pitted by solution once extended around Amantes Point (Pl. 3, fig. 1). At the point itself the 5-foot bench has been completely eroded away and in its place is the shore platform of the present sea which rises from 1 foot above sea level 550 feet northeast of the point to 8 feet above sea level at the point, clearly showing the effect of the waves piling up as they sweep around the point (Fig. 2 and Pl. 3, fig. 1). The platform drops to altitudes only 2 feet above sea level in a distance of 100 feet to the leeward of Amantes Point (Pl. 3, fig. 2). The frontal reef flat in the same stretch declines in level by 1.5 feet so that it is below sea level at the point, but on the windward side the frontal reef rises to meet the shore platform. The depth of scour increases with the height of the waves on the reef flat but on the shore bench the height of the notch increases with the increase in height of the breaker splash. The notch reaches a height of 11 feet in places because of the coalescence of the present sea notch with that of the 5-foot sea. Such huge notches are usually indicative of more than one stand of the sea.

The fact that the present sea has not destroyed the 5-foot bench in many places indicates that the halt at the 5-foot level was probably longer than the present stand.

FIVE-FOOT STAND OF THE SEA

Benches of the 5-foot stand exist on most islands but are readily separated from benches of the present sea only where both occur or where emerged reefs can be found, as both benches may have the same altitude.

The exact extent of the shift, as determined by the difference between notches in a quiet lagoon in Guam, amounts to 5.2 feet (Pl. 3, fig. 3); the emerged atoll on Midway indicates a similar drop in sea level (Pl. 1, fig. 2). A bench about 5 feet above sea level planed across a lithified dune above the present sea platform on Kapapa Islet off Oahu indicates a similar drop in Hawaii. The 5-foot bench is readily distinguished from the present sea platform because it can be walked with safety during high tide, although it is not safe during storms. The fact that land vegetation grows on it in places also indicates that it is unrelated to present sea level. Most of the shore fishing on Pacific Islands is done from the bench of the 5-foot stand, and miles of cliffed coasts would be inaccessible were it not present. As pointed out above, it is true that the 5-foot bench has been reduced in places by weathering and erosion to coincide with the present shore platform, but on exposed headlands it reaches heights of about 15 feet. This bench was formed in the same way as the present sea platform and lay above the sea level of its time.

TWENTY-FIVE-FOOT STAND OF THE SEA

The 25-foot stand of the sea occurred so long ago, probably in late Pleistocene, that its shore platform has been destroyed in most places by subsequent erosion and weathering. However, it can still be found unmodified in protected places and at such sites ranges in altitude from 20 to 35 feet. This stand of the sea lasted long enough for fringing reefs and extensive terraces to be formed. On headlands beach deposits of this sea may reach 35 to 40 feet above sea level, as on Ulupau Head (Stearns, 1935a, p. 122; Wentworth and Hoffmeister, 1939, p. 1565).

CONCLUSIONS

Three shore benches exist in the Pacific. The lowest, made by abrasion and solution of the present sea, ranges from near sea level to about 12 feet above sea level. It is always awash at high tide. In most places the two upper benches are reduced below their original height by subsequent weathering and erosion. The middle one ranges from 3 to 15 feet above

sea level and was formed by a sea about 5 feet higher than the present. It is awash at high tide only where it has been reduced by subsequent weathering and erosion. The upper one is 20 to 35 feet above sea level and was made when the sea stood 22 and 27 feet above its present level. These shore lines are so universal in the Pacific that they must be eustatic, probably indicating shifts in ocean level induced by a late advance in the ice caps and by concurrent changes in the configuration of the ocean floors. The 5-foot recession is very recent; it may have occurred only about 5,000 years ago.

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FIGURE 1. BRECCIA COMPOSED OF LARGE ANGULAR BLOCKS OF ANDESITE
In a matrix of small blocks and tuff.

FIGURE 2. CONGLOMERATE COMPOSED ENTIRELY OF ANDESITE DEBRIS

CHARACTER OF LOWER PLIOCENE ANDESITE

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LATE TERTIARY GEOLOGY SOUTHEAST OF MONO LAKE, CALIFORNIA

BY CHARLES M. GILBERT

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ABSTRACT

The region southeast of Mono Lake is more or less representative of much of the western Great Basin. Pliocene volcanic rocks lie on an old erosion surface cut in pre-Tertiary granitic and metamorphic rocks. The oldest are rhyolite tuff and flows overlain in turn by andesite flows, tuffs, and breccias and by olivine basalt flows. Small erosional breaks separate these three series. In the central part of the area, a mountainous mass of rhyolite overlies the basalt, and eastward streams have spread a mantle of rhyolite tuff and gravel.

All formations are displaced by normal faults, along the largest of which the present ranges have been relatively elevated. Although this faulting has continued until the present, it was nearly complete by middle(?) Pleistocene time. The result is a mosaic of fault-bounded blocks, and statistical correlation of the strikes of these faults and older joints supports the view, suggested by others, that the individual faults are controlled by earlier planes of weakness, notably by joints in the granitic rocks. The factors controlling the location and trend of the major displacements which produced the ranges are unknown.

Volcanic activity, following the initiation of the Basin-Range faulting, continued through the Pleistocene until recent times.

INTRODUCTION

Immediately southeast of Mono Lake, the Sierra Nevada on the west and the White Mountains on the east bound a large complex graben about 35 miles wide which includes several smaller fault-block ranges. Within this graben lie late Tertiary and Pleistocene volcanic rocks, both younger and older than the Basin-Range faulting which produced the graben. This region is more or less representative of other parts of the western Great Basin both in structure and in exposed rock types. This paper describes the late Tertiary volcanic rocks and their relationships, suggests a correlation between them and similar volcanic rocks in adjacent areas, and discusses the Basin-Range faults within the area.

ACKNOWLEDGMENTS

The writer is indebted to the members of the Department of Geological Sciences of the University of California at Berkeley and to V. L. VanderHoof and R. A. Stirton of the Department of Paleontology for suggestions and criticism; to E. S. Shepherd of the Geophysical Laboratory in Washington for permission to use several analyses; and to H. G. Ferguson,

W. H. Swayne, and F. E. Matthes for communications concerning their work in near-by areas.

OLDER FORMATIONS

The oldest rocks exposed in this area are all metamorphosed sediments—marble, mica schists, quartzite and calc-silicate hornfels being

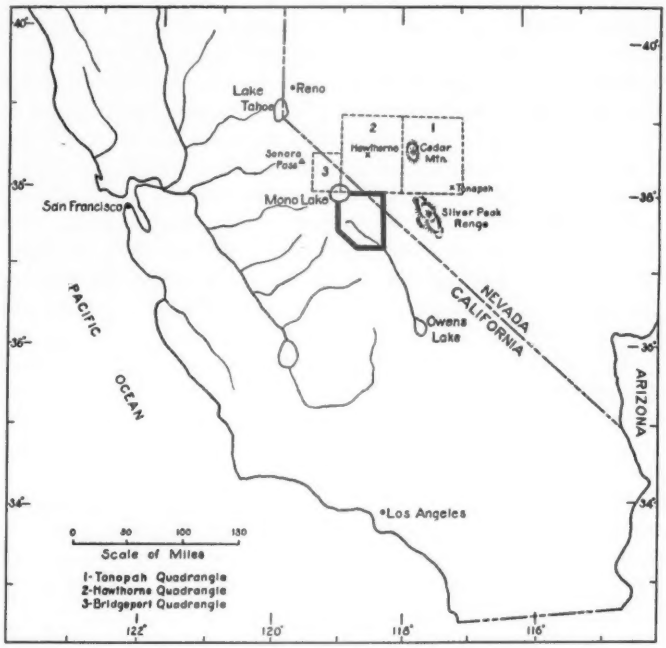

FIGURE 1.—*Sketch map of California*
Showing location of area studied and localities to which reference is made.

the most common. These are intruded by granitic rocks so widely exposed by erosion that the metasediments are numerous small patches of obscure structure. Universal steep dips prove that they have been intensely folded.

Among the granitic rocks are several different types and a number of separate intrusions. Most common are porphyritic granodiorite with large phenocrysts of orthoclase and an even-grained type grading into the porphyritic variety wherever the two are in contact. Several small intrusions of quartz monzonite, deeply weathered and nearly free from

ferromagnesian minerals, are exposed, one in the Benton Range extending several miles southward from Black Lake and others on Casa Diablo Mountain and near Granite and Black Mountains, in addition to a few small masses of hornblende diorite. Presumably these plutonic rocks are related to the Sierra Nevada batholith which is regarded as late Jurassic.

Intrusive into both plutonic and metamorphic rocks is a swarm of thick granite or rhyolite porphyry dikes trending consistently N 5°–15° W in the Benton Range from Casa Diablo Mountain northward for 5 or 6 miles. Prominent phenocrysts of quartz, orthoclase, perthite, and occasional flakes of biotite are embedded in a microcrystalline groundmass in the larger dikes. Small dikes and the chilled margins of larger ones are cryptocrystalline. This dike swarm intrudes all the older rocks except the quartz monzonite, though several occur within it. Probably the quartz monzonite was more impenetrable, forming a sort of barrier to intrusion perhaps because of its more widely spaced joints. Later than these acid dikes are some transecting malchite dikes. Consisting mainly of commonly altered hornblende and plagioclase, these malchite dikes are fine-grained and some contain both feldspar and hornblende phenocrysts. None parallels the acid dikes; all strike either northwest or west as though crowding in an east-west direction, resulting from the intrusion of the northward-trending dike swarm, prevented any parallel intrusion of the later malchites.

The age of these dikes is not directly evidenced. They were eroded before the outpouring of the late Tertiary volcanic rocks. They are younger than the Jurassic plutonic rocks which they intrude, and their chilled margins indicate intrusions between cool walls, possibly long after consolidation of the plutonic rocks. Nearly all the gold and silver prospects and ore deposits of the region lie in the area occupied by these dikes, a fact which suggests a genetic relationship.

LATE TERTIARY VOLCANIC ROCKS

OLDER RHYOLITE

From the stratigraphic standpoint the most important occurrence of the older Tertiary rhyolite is on the summit and east flank of the Benton Range about 4 miles north-northwest of Benton. It had been somewhat eroded before the overlying series of andesite breccias, conglomerates, and flows was emplaced. Within half a mile from north to south along the crest of the range it is gone and the overlapping andesite lies on the pre-Tertiary rocks to the south.

Composed entirely of vitric-crystal tuff, the rhyolite in this locality is divisible into three massive members on the basis of color and content of crystals and lapilli. The lowest member is exposed at the top of the

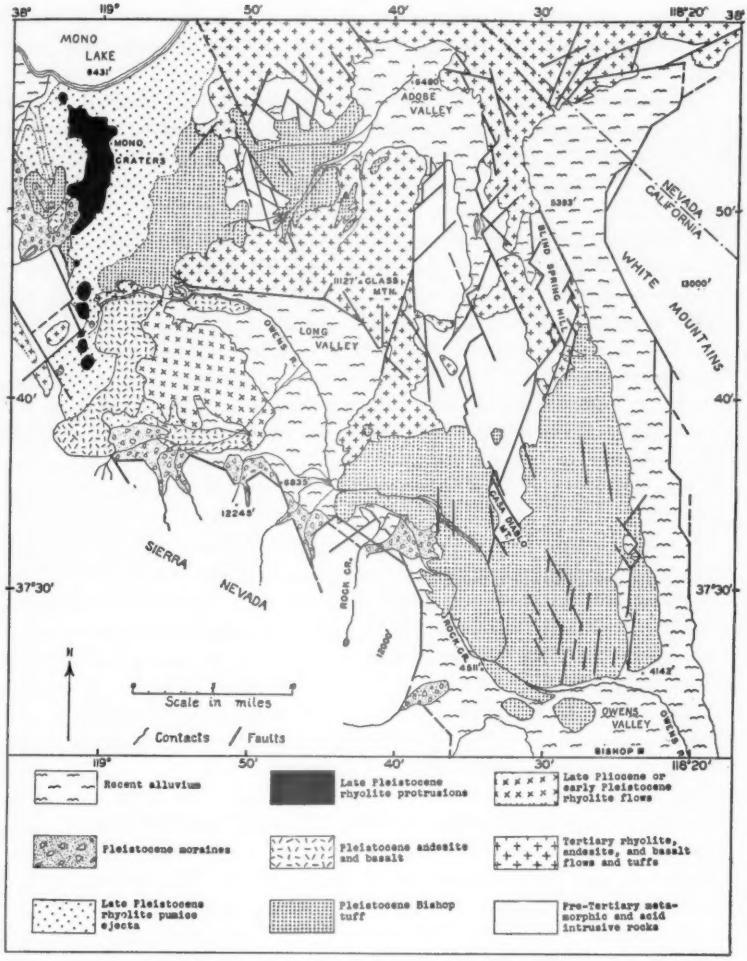

FIGURE 2.—*Geologic map showing distribution of Pleistocene volcanic rocks*

"Late Pliocene or early Pleistocene rhyolite flows" in the legend are probably middle or late Pleistocene in age according to Chelikowsky (1940, p. 423) who has studied them in more detail.

fault scarp which forms the west side of the Benton Range. It is about 25 feet thick, has no trace of bedding, and is composed of a dull-gray matrix of poorly sorted and partially devitrified glass shards in which are scattered crystals of quartz, sanidine, larger fragments of light- and

dark-gray pumice, and pellets of obsidian. The refractive index of the glass varies slightly about 1.490, and the crystals compose only 5 per cent of the whole.

The middle member is thicker. More abundant fragments of biotite, sanidine and quartz (quartz predominant) are scattered in a dull-white matrix presumably glass dust but now so devitrified as to appear homogeneous. Sparse lapilli of white devitrified pumice are also scattered through it.

In contrast, the uppermost member of the rhyolite tuff is dark gray and differs from those below in its greater content (about 35 per cent) of crystals of quartz, sanidine, biotite, basic oligoclase, and hornblende and by the presence of lapilli of porphyritic pumice up to 2 inches in diameter. Sanidine is most abundant, and nearly euhedral crystals are common; quartz and biotite are next in abundance; plagioclase and hornblende are sparse. Residual glass in the matrix and in the pumice has a refractive index of 1.500 ±.003. Like the basal member, this one is about 25 feet thick, but in places it has been removed by erosion prior to the outpouring of the more basic volcanics and may nowhere represent its original thickness.

It thus appears that the tuff becomes progressively richer in crystals toward the top and that more mineral species occur where the crystals are more abundant. These features suggest that the tuff is a subaerial deposit and that the material was supplied by volcanic explosions rather than by erosion of pre-existing rhyolitic rocks, the successive tuff members representing eruptions of magma at successively later stages of crystallization. That no abrasive or sorting agent was important during deposition is further indicated by the perfection of some of the sanidine crystals and by the absence of bedding.

In the hills just northwest of Queen, Nevada, are flows of cryptocrystalline tridymite rhyolite and spherulitic and perlitic obsidian. The relation between these and the andesite flows, breccias, and conglomerates with which they are in contact is not entirely clear. The rhyolite appears to lie below the andesite and may be the same age as the rhyolite tuff in the Benton Range, possibly the source of that tuff. However, this rhyolite may intrude the basal part of the andesite series. Certainly its contacts with the andesite are irregular and rather steep. Either of two possible explanations seems plausible: the rhyolite may be a protrusion, irregular and steep-sided, which was overwhelmed by the andesites; or it may represent a protrusion which rose through the lower part of the andesite series. In either case the extrusions of rhyolite and andesite are separated by no important interval of time; indeed, the periods of their extrusion may overlap.

ANDESITE

General character.—Although this andesite series is thicker and more widely distributed to the northeast, its stratigraphic relations are clearest

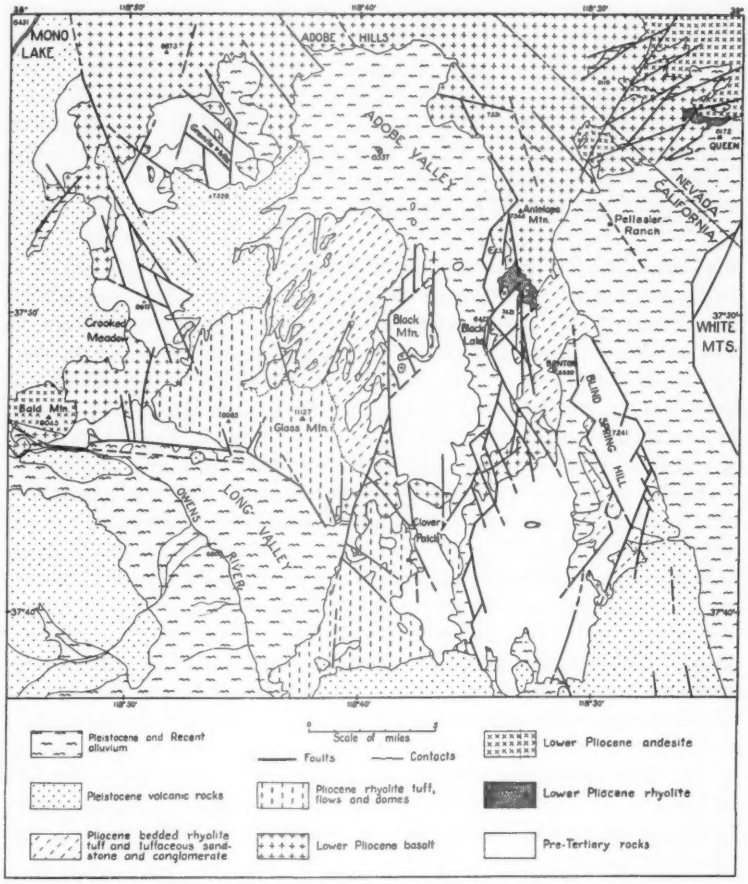

FIGURE 3.—*Geologic map showing distribution of Tertiary volcanic rocks*

at the southernmost occurrence on the eastern flank of the Benton Range. It dips about 20 degrees east, separated by slight erosional unconformities from rhyolite tuff below and olivine basalt flows above. The maximum thickness of the andesite is about 200 feet in this locality. Probably

it was never much thicker and never extended much farther south. It occurs as remnants isolated by Pleistocene and recent erosion consequent upon the uplift of the Benton Range.

Massive beds of breccia containing blocks of hornblende-pyroxene andesite up to 3 feet in diameter make up most of the andesite series in this locality (Pl. 1, fig. 1). Interbedded are one flow of hornblende-pyroxene andesite and beds of andesitic conglomerate and tuff. The beds of conglomerate range in thickness between 1 and 10 feet and contain only andesitic material of which the larger fragments are subangular and subrounded (Pl. 1, fig. 2). The finer tuff members are fairly well sorted, and the bedding is well defined.

North-northeast from this locality, a broad expanse of olivine basalt covers the andesite for several miles. Beyond this basalt 600 feet of andesite is exposed on the range scarp north of Pellesier Ranch. Here, as to the south, flows are subordinate to breccia, but two features are noteworthy: (1) In the conglomerate which composes the uppermost 100 feet of the section a few boulders of granodiorite, quartzite, and andesite tuff occur with those of andesite; (2) a layer of andesite tuff near the base of the exposed section contains fragments of white rhyolite pumice.

In the hills northwest of Queen, Nevada, the andesite series contains thinner and less numerous conglomerate and tuff beds; flows and breccia dominate the section (500 feet thick) overlain by olivine basalt. Farther west welded andesite tuff, exposed on several fault scarps, lies beneath the olivine basalt flows with no apparent discordance and seems to represent the final phase of the andesite eruptions. The tuff is a hard rock containing many small lenses of dull-black glass lying nearly horizontal. Abundant crystals of plagioclase and biotite, a few of pyroxene, and fragments of biotite-pyroxene andesite up to 4 inches in diameter are scattered through it. The formation is massive and seems to be composed of layers 35 to 50 feet thick, best seen from a distance, dipping gently westward. Close at hand there is no trace of layering in the tuff except as revealed by alinement of the glass lenses. Insufficient data are at hand to permit a discussion of the origin of this tuff, but it is strikingly similar to the welded Bishop tuff (Gilbert, 1938, p. 1829-1862), exposed to the south.

Petrography.—The andesites in the flows, breccias, and conglomerates are essentially alike. All are porphyritic with pilotaxitic, hyalopilitic, or occasionally glassy groundmasses. Phenocrysts of plagioclase show oscillatory zones varying between sodic labradorite and medium andesine. Most of these are euhedral with a thick tabular habit parallel to the side pinacoid (010); some reach a length of 3 millimeters and many contain

abundant inclusions of pyroxene, magnetite, apatite, and clear glass. Less common and smaller phenocrysts are augite and hypersthene and of these augite is the larger and more abundant. In addition, phenocrysts of partially resorbed basaltic hornblende are present in some specimens. More notable are the crystals of apatite some of which are large enough to be classed as phenocrysts; as many as a dozen prisms half a millimeter long occur in some slides.

The groundmass of the typical specimen is a fine felt of medium andesine laths with interstitial augite, magnetite, and glass. Some specimens contain no glass but most have as much as 5 or 10 per cent. Magnetite is both euhedral and anhedral; some is included in the phenocrysts, but much of it crystallized with the groundmass constituents, and where glass is abundant it is so clouded by magnetite dust as to be almost opaque.

Origin.—Observations suggest that the sources of the andesites were central vents around which cones may have risen. Pointing toward such a view is the rather rapid thickening of the series to the north, and the variation in the series from one place to another. Whereas flows may be subordinate at one locality, they are abundant at another, seemingly more numerous northward. The breccias suggest mudflows as the agent of deposition, as on the flanks of some large modern cones, while the conglomerates and tuffs containing little but andesite debris were perhaps deposited mainly on the lower slopes of the volcanoes by streams and floods. Where the vents lay is unknown, but from the thickening of the series, the occurrence of more flows, and the culminating welded tuff to the north probably they are to be found in that direction, perhaps beyond the limits of the map.

OLIVINE BASALT

General character.—The most widespread Tertiary formation is a series of noncontinuous olivine basalt flows which are encountered throughout the mapped area. Although some basalt cappings on older rocks are relatively thin, these flows accumulated to a thickness of 600–700 feet east of Bald Mountain and to the west on the Sierra Nevada crest. Only olivine basalt is exposed in the Adobe Hills where the series must be thicker since moderately dipping flows cover an area of considerable relief.

Most flows vary between 25 and 50 feet in thickness and may be easily distinguished, one from another, by the scoriaceous zone at the top of each and commonly by a basal vesicular zone. In only two localities, Owens River Gorge beneath the Bishop Tuff and the Sierra Nevada crest west of Bald Mountain, does fragmental material occur between the flows; there the flows are thinner and clearly defined by intercalation of easily eroded beds of basalt breccia.

In the Benton Range, north-northwest of Benton, the basalt rests on both the andesite and rhyolite tuff. Prior to the eruption of the basalt erosion removed the andesite from some places, so that the basalt lies directly on the older rhyolite tuff. Although these older volcanic rocks, and especially the andesite, are widespread beneath the basalt to the northeast, they are not encountered in any other portion of the mapped area suggesting that the andesite and rhyolite accumulations were more localized.

Petrography.—The usual basalt is light gray and porphyritic with intergranular groundmass; only a few flows are black and glassy. The phenocrysts are small crystals of olivine, augite, and plagioclase (An_{65}–An_{75}). Colorless olivine phenocrysts, mainly unaltered, range in diameter up to 2 millimeters, and the optic axial angle (2V) of 90 degrees shows that the olivine is magnesia-rich. Augite phenocrysts are prisms 1 millimeter long with a greenish tinge under the microscope, strong dispersion, and a common hourglass structure. Plagioclase, the least abundant of the phenocrysts, varies between calcic labradorite and sodic bytownite; it is rarely zoned and occurs as small thick tabular crystals. Ten to 15 per cent of average specimens are phenocrysts and half to two thirds of these are olivine.

Laths of labradorite with intergranular augite and magnetite constitute most of the groundmass; some specimens have a few very small olivine grains which may belong to the groundmass generation. Although some of the magnetite—that having euhedral form and included in phenocrysts—crystallized before other groundmass constituents much must have crystallized later since it occurs as subhedral or anhedral grains in the groundmass. Indeed, in glassy specimens, magnetite dust makes the glass nearly opaque.

A crystalline mesostasis is the only unusual feature of these basalts. Quantitatively unimportant in most specimens, it consists of minute untwinned feldspar crystals crowded with apatite needles and associated with flakes of brown phlogopitic biotite. It is always xenomorphic and moulded around the other constituents. Ransome (1909, p. 73-74) mentioning such constituents in the Tertiary basalts of the western Great Basin states that at Goldfield the Malpais basalt contains, in places, a little orthoclase and biotite as final crystallization products. In contrast, the untwinned feldspar in the basalts near Mono Lake is sodic since the refractive index for the fast ray is $1.530 \pm .002$. The biotite probably contains considerable magnesia although it is not true phlogopite: *Z*—reddish-brown, *X* and *Y*—yellow; *Z*—$1.611 \pm .003$; dispersion extreme; optic axial angle (2V) as large as 30 degrees but variable. The biotite and sodic feldspar apparently formed contemporaneously since they never

occur separately. Some vesicles are lined with small crystals of albite and flakes of lustrous red-brown biotite.

This mesostasis forms less than 1 per cent of some specimens, but in others it constitutes 15 or 20 per cent. Wherever significant quantitatively the abundance of the mesostasis seems to stand in inverse proportion to the abundance of labradorite, due to a partial reaction between the two. Some labradorite shows only a thin border of reaction, while elsewhere it loses its clear outline, grading into the sodic feldspar. Where the latter occurs the refractive index is highest near the central core of the original labradorite lath gradually decreasing toward the margin; a progressive change of extinction angle outward from the center occurs, and apatite needles which are absent in the center increase in abundance outward. The sodic feldspar thus becomes more calcic though variable, while the labradorite lath, a mere shadow of its original form, becomes more sodic along its margins. Where this reaction has not occurred the mesostasis is quantitatively unimportant, and the untwinned feldspar composing it has the refractive indices of albite and contacts the labradorite along sharp boundaries. Were a perfect reaction to occur between the sodic feldspar in the mesostasis and the labradorite in the average sample the resulting feldspar would probably approximate sodic labradorite or calcic andesine in composition, though it is difficult to determine an average since the quantitative significance of the mesostasis varies greatly from one flow to another and within any one it increases toward the center. The striking thing about the mesostasis is its frequent occurrence in these basalts.

The analysis (Table 1, No. 1) by Dr. E. S. Shepherd represents an unusual specimen from the center of a thick flow at the summit of the Benton Range about 5 miles south of Black Lake. White patches the size of a dime, representing concentrations of the mesostasis with a little interstitial analcite, splotch the usual gray groundmass. Abundant large flakes of brown biotite occur in the vesicles. This specimen represents abnormalities greater than those found in any other flow. Unusually high for a basalt in K_2O, P_2O_5, and F, it also contains more MgO and less Al_2O_3 than most. The biotite is insufficient to account for the potash present; some is probably in solid solution in the plagioclase, perhaps in the mesostasis. Nor does the apatite content of the rock seem sufficient to account for the amount of P_2O_5 shown in the analysis.

Whatever the genetic significance of these findings may be, these conclusions are justified:

(1) The apatite, biotite, and sodic feldspar were the last constituents to crystallize in the basalts and are genetically related.

(2) The mesostasis represents a residuum in a crystallizing magma, and is composed of constituents present in the magma at the time of extrusion. Had the mesostasis been formed by introduction of material

TABLE 1.—*Analyses and norms*

Analyses

	1	*2*	*3*
SiO_2	49.97	76.27	76.13
Al_2O_3	12.83	12.24	12.49
Fe_2O_3	3.84	.50	.36
FeO	4.95	.76	.72
MgO	10.20	.02	.02
CaO	8.48	.60	.54
Na_2O	3.49	4.90	4.68
K_2O	2.11	4.49	4.72
H_2O+	.43	.15	.20
H_2O-	.15	.06	.09
CO_2	.00	.....	.....
TiO_2	1.42	.10	.08
ZrO_2	?	.00	.00
P_2O_5	.80	.04	.02
Cl	.01	.11	.13
F	.10	.06	.07
S	.03	.01	.....
Cr_2O_3	.08	.....	.....
MnO	.04	.01	.02
BaO	.19	.00	.00
Li_2O	.017	.....	.....
Total	99.137	100.32	100.27

Norms

Quartz	.....	31.02	30.66
Orthoclase	12.23	26.69	27.80
Albite	29.34	37.73	38.25
Anorthite	13.07	.....	.....
Olivine	13.05	.....	.....
Diopside	17.97	.....	.....
Acmite	.....	1.39	.46
Na metasilicate	.....	.30	.....
Wollastonite	.....	1.10	.93
Hypersthene	1.76	1.32	1.06
Magnetite	5.57	.....	.....
Ilmenite	2.74	.15	.15
Apatite	2.02	.....	.....
Fluorite	.25	.12	.16
Na chloride	.....	.18	.23

[1] Olivine basalt from center of thick flow at top of Benton Range fault scarp south of Black Lake.
[2] Obsidian collected about 1 mile northwest of Glass Mountain.
[3] Obsidian from end of northeast flow, Mono Craters.
(All analyses by E. S. Shepherd, Geophysical Laboratory, Washington, D. C.)

subsequent to crystallization, it would have altered the other constituents and would occur mainly near eruptive vents, having no variation in importance with each flow.

Attention must also be called to flows of andesitic character here and there in the basalt series. Near the Clover Patch are several hornblende-

bearing flows, identical with the usual basalt except for large resorbed phenocrysts of hornblende replaced by aggregates of fine magnetite and augite. Their greater size suggests that hornblende started crystallization early, explained by assuming that a considerable volatile content in the magma favored crystallization of hornblende in the intratelluric stage under pressure. The presence of volatiles is also indicated by the composition of the mesostasis.

Occasional flows are true andesites containing hornblende and pyroxene in pilotaxitic or hyalopilitic groundmasses. Petrographically indistinguishable from the earlier andesites, in the field they are identified by their association with the basalt and by the absence of fragmental accumulations so important in the Tertiary andesite series.

Structural relations and origin.—The olivine basalt was erupted before fault displacements formed the present ranges, for the flows, as gently tilted lava fields elevated by faulting, cap many of the ranges. This sequence may be observed at Granite Mountain, in the Benton Range south of Black Lake, in Owens River Gorge, and on the plateaulike summits between Bald and Glass Mountains.

The erosion surface buried by the basalt series had little relief. Mantled as it is by basalt and displaced by faulting, this old surface cannot be seen continuously for more than 1 or 2 miles in any one place, but it was not a plane surface since parts of it, including several granitic hills of considerable height, were never buried by the basalt. Several miles north of Benton bedded rhyolite tuffs, faulted and slightly younger than the basalt, overlie the basalt flows, while to the south near Benton these tuffs lie on a deeply weathered surface of slight relief cut in granitic rocks. The basalt, had it ever mantled this part of the old surface, could scarcely have been removed in the short period preceding the deposition of the tuffs, certainly not without dissection of the older surface.

Several basaltic vents are known but many more must have existed. Two necks stand near Crooked Meadow, and several smaller ones northwest of Granite Mountain lie in a line half a mile long trending north-northwest. Two scoria-covered hills north of Granite Mountain and another southwest of Blind Spring Hill may also mark the site of vents, and the basaltic peak north of Clover Patch probably represents another. The flows with intercalated basaltic breccias exposed in the gorge of Owens River and those on the crest of the Sierra Nevada indicate other centers since the breccias are peculiar to those places.

Apparently a deeply weathered old-age erosion surface was flooded, chiefly in the broad valleys, by basaltic flows which issued from scattered vents. An unbroken basalt field across the entire area never existed

for the higher parts of the old surface rose above the lava. Basaltic remnants are distributed on the lower parts of the crest of the Sierra Nevada lying in ancient broad valleys now uplifted. Indeed a small basalt exposure on McGee Mountain lies in an old valley, later the path of an early glacier which left its moraine on the basalt (McGee Stage of Blackwelder, 1931, p. 902-906).

ANDESITE OF BALD MOUNTAIN

The hornblende andesite on the summit and western side of Bald Mountain differs from other andesites in the region and must therefore be considered separately. On the summit the andesite is generally reddish and granular because of oxidation of iron, though occasional small masses of gray unaltered hornblende andesite occur. On the southwestern slope, however, a considerable thickness is exposed, and the oxidation decreases downward into a perfectly fresh dark gray andesite. A platy structure to which hornblende prisms are roughly parallel probably represents flow planes. This dips into the mountain at angles of about 20 or 25 degrees.

The field relations of this andesite cannot be interpreted with certainty. On the east side of the mountain it meets plutonic rocks and Tertiary basalt along a presumably intrusive contact dipping steeply and striking irregularly; its southern contact with basalt is likewise intrusive but more regular in strike. Elsewhere it is overlain by the Pleistocene Bishop tuff.

Assuming the mass to be intrusive, it is younger than the olivine basalt. Its relation to the Glass Mountain rhyolite cannot be determined. However the intrusion must have antedated the Basin-Range faulting for the south side of Bald Mountain is a scarcely dissected fault scarp nearly 2000 feet high and the andesite is separated from the scarp by only a few hundred feet of basalt.

Fresh andesite from the lower part of the protrusion contains numerous small phenocrysts of green hornblende and calcic andesine in a hyalopilitic groundmass containing 5 to 10 per cent of glass. The remainder of the groundmass consists of minute laths of andesine, tiny prisms of augite, and flecks of magnetite. Apatite in prisms up to 0.2 millimeter in length is common but is undoubtedly of an earlier generation than the groundmass since it is included in the phenocrysts.

Higher in the protrusion the only notable change is in the hornblende which is replaced almost totally by granular magnetite and augite. In the red, oxidized andesite, on the summit, the hornblende is completely replaced by hematite and augite. Furthermore, the rare spots of unresorbed hornblende are the basaltic type with pleochroism in reddish-brown and yellow and a very small extinction angle. In this oxidized andesite occasional crystals of quartz show extremely irregular outlines

and include tiny prisms of augite like those in the groundmass. Consequently, the author interprets them as products formed during the late stage of solidification along with the oxidation.

This petrographic picture accords well with the idea that a shallow intrusion broke through to the surface as a domelike protrusion. By relief of pressure consequent upon rise of the magma to the surface, hornblende was altered to basaltic hornblende and resorbed. The oxidation and deposition of silica was probably accomplished by gases escaping from and through the hot dome and decreased with depth in the crystallizing body.

LATER RHYOLITE

Glass Mountain area.—The slopes of Glass Mountain and neighboring peaks are mantled by rhyolite debris. Outcrops of the flows which compose the range reveal a variety of both glassy and lithoidal rhyolitic types. Possibly beds of tuff exist between the flows though none is clearly exposed.

On the northern slopes of Glass Mountain and the peaks adjacent the flows have a general low dip defined by the flow banding. This dip is not constant even within one flow, but the generally low angle indicates true flows. On the summits and along the southern slopes, uniformly steep dips prevail and there are many small plugs of obsidian, suggesting that the vents from which the rhyolite was extruded lay along the summit and southern flank of the range. (One vent beneath Glass Mountain is inferred in Figure 4, *B*.) The range was probably built by a series of large domelike protrusions from which short thick flows occasionally broke. Subsequently this rhyolite, several thousand feet thick near Glass Mountain, was displaced by faulting so that its southern part must now lie beneath the alluvium in Long Valley.

Whether the faults along the southern scarp of Glass Mountain were controlled by the rhyolitic vents is not known, but clearly the faulting succeeded the rhyolitic eruptions for the southern slope of the range is a fault scarp (Pl. 2, fig. 2) along which not only the rhyolite but the underlying granitic and basaltic rocks east and west of Glass Mountain have been displaced.

The central rhyolite mass lies in a graben in the older rocks (Fig. 4, *B*). At the north base of Glass Mountain the basalt and pre-Tertiary granitic rocks lie between 1000 and 1500 feet higher than at the southern base along the fault scarp. To the west the granitic surface beneath the rhyolite slopes downward toward Glass Mountain, and eastward it rises again, elevated by faulting. Thus the basement of Glass Mountain is an irregular, basinlike, structural depression, perhaps formed by downwarp-

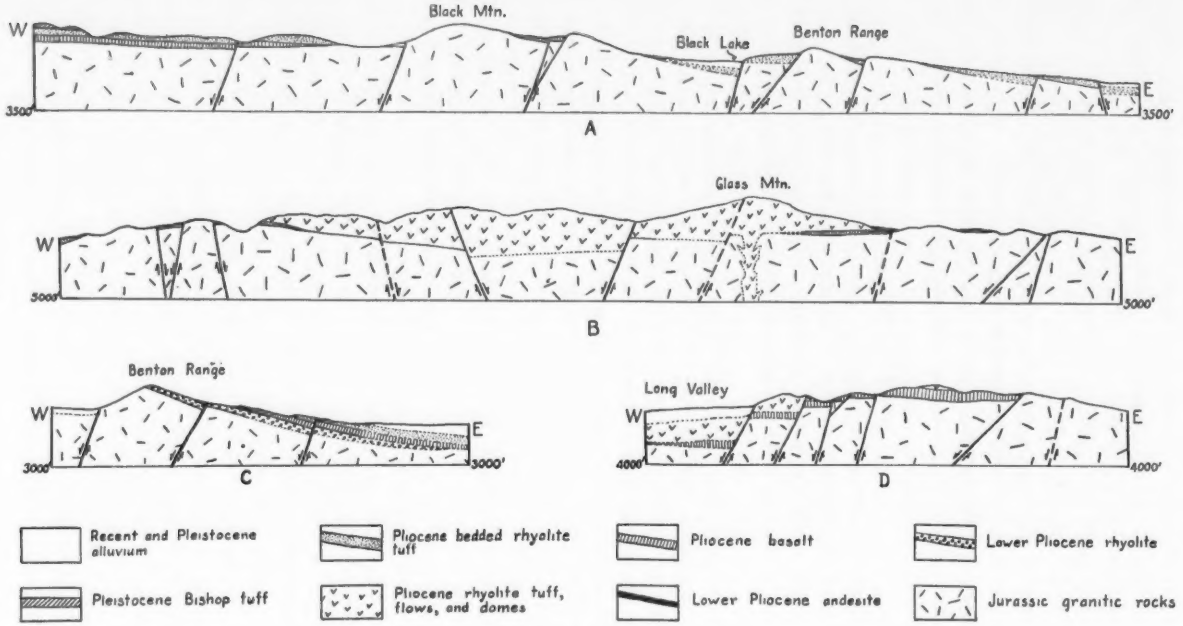

FIGURE 4.—*Geologic structure sections*

Vertical and horizontal scale—1 inch equals 2 miles. A—east-west section through center of Black Lake; B—east-west section through summit of Glass Mountain; C—east-west section across Benton Range 3½ miles north of Benton; D—east-west section just north of Clover Patch.

ing and faulting during the rhyolite eruptions, perhaps after these eruptions, or possibly both during and after the eruptions.

The lithoidal rhyolite composing the flows north of Glass Mountain contains phenocrysts of quartz, sanidine, and oligoclase in a granophyric groundmass. Occasional phenocrysts of biotite are reddened and partially converted to granular magnetite. In contrast, the rhyolite on the summit and southern scarp is characterized by abundant tridymite and by cryptocrystalline and spherulitic texture. The granophyric texture is conspicuously absent, and tridymite composes nearly one third of some specimens, suggesting the proximity of the main vents.

The analyses of obsidians (Table 1, Nos. 2 and 3), one collected about 1 mile northwest of Glass Mountain and the other from the northeast flow of the Mono Craters, were made by Dr. E. S. Shepherd. They are strikingly similar considering their difference in age and location, both being notably high in soda and silica and low in alumina.

Clover Patch area.—On the hills around Clover Patch, southeast of Glass Mountain (Fig. 3), and extending south until it disappears beneath the Bishop tuff, is a mantle of rhyolite tuff overlying the olivine basalt and displaced by the normal faults. In the tuff, fragments of brilliant white pumice, chips of lithoidal rhyolite like that on Glass Mountain, pellets of obsidian, and crystals of quartz and feldspar are scattered through a matrix of glass dust. Here and there on the tuff-mantled area are conical hillocks about 50 feet high where blocks of lithoidal rhyolite up to 2 feet in diameter are concentrated. Since two small rhyolite pipes drilled through the granodiorite are exposed just east of Clover Patch, these hillocks are very likely similar vents from which the tuff mantle has not yet been stripped. Thus, to the east of the main eruptive centers near Glass Mountain there was a field of small eccentric pipes which erupted explosively blanketing the surrounding area with rhyolite tuff. The Glass Mountain protrusions and the minor pipes near Clover Patch, with their similar lithology, stratigraphic relations, and their proximity, must be related.

Bedded rhyolite tuff.—Sloping gently northward toward Adobe Valley from the base of Glass Mountain is a dissected plane underlain by unconsolidated rhyolite tuff and gravel. The tuff is composed of small fragments of glass, quartz, feldspar, and abundant subangular pebbles and small boulders of rhyolite and pitted obsidian, undoubtedly derived from Glass Mountain and its neighboring peaks and spread by streams to the north and east. However, the nature of the material beneath the surface is obscured by debris mantling the canyon walls.

These tuffs overlie the olivine basalt flows exposed in some of the canyon bottoms, and they are in turn overlain by the Bishop tuff which forms a more resistant brown capping. Several normal faults displace them, notably along the west flank of Black Mountain and west-northwest of Glass Mountain.

On the eastern flank of the Benton Range for several miles north and south of Benton, similar tuffs are widespread, lying on the Tertiary basalt, or on the granitic rocks where the basalt is absent (Pl. 3, fig. 2). They differ from the tuffs north of Glass Mountain only by the presence of a few pebbles and small boulders of granitic rocks, olivine basalt, quartz porphyry, and metamorphic rocks. In the sides of many canyons and gullies these tuffs are well exposed, dipping gently eastward a little less steeply than the surface upon which they rest and displaced by a number of normal faults (Pl. 3, fig. 2).

Evidently these bedded tuffs in the Benton Range were once connected with those north of Glass Mountain for the general character, composition, and relations of the tuff in the two areas are similar. A remnant of the same tuff at the foot of the Benton Range fault scarp east of Black Lake about midway between the two more extensive tuff areas suggests an earlier connection between them. The once continuous tuff blanket has been dismembered by faulting and the consequent stripping of the tuff from the range summits (Fig. 4, *A*). This interpretation is at variance with that of Ransome (1940, p. 168-170) who states that the tuff is older than the Tertiary basalt and is not faulted.

The following field characteristics of the formation in the Benton Range are readily determined:

(1) The basal beds of the series are tuffaceous conglomerates containing subangular to rounded boulders of the bedrock beneath.

(2) The upper part of the series is entirely rhyolitic. Light-colored pumice, lithoidal rhyolite, and obsidian make up the larger fragments which include both angular chips and subangular to subrounded pebbles and cobbles, while the finer fragments are glass, quartz, and feldspar. Most beds are a poorly sorted aggregate of all of these fragments, but a few contain only pumice and some consist entirely of small grains of glass and crystals.

(3) The formation is massive and cross-bedded. While most beds are at least 2 feet thick and some as much as 10 feet thick, occasional beds are only a few inches in thickness.

(4) The maximum exposed thickness of the series near Benton is about 200 feet, and a similar thickness is exposed near the south end of Blind Spring Hill. The total thickness probably does not exceed 500 feet.

Probably stream deposits, these massive beds suggest rapid deposition. The author pictures a surface of low relief, drained eastward by either one broad stream valley or by a number of smaller streams. A veneer of gravel deposited by the streams was mixed with rhyolitic material derived from the early eruptions near Glass Mountain. From time to time, when the vents around Glass Mountain were active, the surrounding country was mantled by ash and the overloaded streams deposited material in thick, poorly sorted beds forming the upper part of the series. As the fault movements began to elevate the present ranges, the unconsolidated tuffs, readily stripped from the higher parts of the rising ranges, were deposited downslope on top of the original beds as "Pleistocene alluvium." Erosion still continues this stripping action.

PLEISTOCENE FORMATIONS

Following the initiation of normal faulting, volcanic eruptions continued through the Pleistocene period until recent times.

The earliest of the Pleistocene volcanic rocks in the region is the Bishop tuff, a welded rhyolite tuff covering approximately 400 square miles between Mono Lake and Bishop (Gilbert, 1938, p. 1829-1861). The Basin-Range faulting was nearly complete when this tuff was deposited by *nuées ardentes* (burning clouds) which issued from scattered vents, mainly near Long Valley, and buried earlier topographic irregularities. The Bishop tuff now forms the "tableland" areas north of Bishop, southeast of the Mono Craters, and west of Adobe Valley.

Subsequently, in Long Valley (Mayo, 1934, p. 95-96) a lake carved terraces on the Bishop tuff and some of the older moraines. A series of Pleistocene sediments containing much volcanic debris was laid down in the lake; now this series projects here and there through the alluvium in the valley.

Whether the basalt flows and andesite domes near the Sierra Nevada scarp to the west preceded or followed the draining of this lake is not clear. Several basalt flows poured eastward into the western part of Long Valley and appear to be overlain in places by domelike protrusions of biotite-hornblende andesite. Another flow of olivine basalt moved eastward from the base of Sierra Nevada toward the Mono Craters, and a small dome of andesite rose near the base of Glass Mountain in Long Valley. These basic volcanic rocks were erupted late in the Pleistocene period, for they are only partly mantled by the small moraines of the last glacial stage.

Also late in the Pleistocene period, viscous rhyolite flows and domes issuing from fissures in the western part of Long Valley produced a

group of ridges and hills (Chelikowsky, 1940, p. 421-436). This rhyolite is younger than the Long Valley lake beds which it intrudes. At about the same time rhyolitic eruptions began to form the line of domes known as the Mono Craters and their southern extension (Russell, 1889, p. 371-390; Mayo, 1936, p. 81-97; Putnam, 1938, p. 68-82). Explosive eruptions accompanying the protrusion of these domes scattered a mantle of pumice over the surrounding region. The latest of the protrusions, the northernmost dome of the Mono Craters (Panum Crater), rose after the high water stage of Mono Lake. Indeed Panum Crater is a perfectly preserved dome encircled by the crater rim of a small pumice cone. Volcanic activity in this region is not entirely a thing of the past for hot springs are still rather numerous in the southwestern part of Long Valley and on Paoha Island in Mono Lake.

CORRELATION AND AGE OF TERTIARY ROCKS

GENERAL STATEMENT

The slightly altered volcanic rocks southeast of Mono Lake, lying with a marked unconformity upon the lake Mesozoic granitic rocks, are Tertiary in age as shown by correlation with other near-by districts. The value of such correlation of volcanic rocks is rather doubtful. For example, if one could unquestionably recognize a single thin tuff or lava flow in two areas, a perfect correlation would be established since the time represented by the formation of a thin tuff or flow is a geological instant. However, in this region, two volcanic series must be correlated upon a comparison of petrography, sequence of petrographic types, and relationships to regional diastrophic events. Such a comparison is a purely descriptive correlation which probably approximates a time correlation (Table 2).

In the western part of the Great Basin, the Tertiary volcanic rocks occur in association with lacustrine and fluviatile sediments. These sediments make correlations more positive although they have not been traced continuously between the regions mapped. Fossils, which are not found everywhere, show that the sediments are not strictly contemporaneous in all localities.

CORRELATION WITH AREA NORTHEAST OF MONO LAKE

The Hawthorne and Tonopah quadrangles, Nevada, (Fig. 1) a large area north and east of Mono Lake, have been mapped by H. G. Ferguson who kindly submitted (personal communication) the following general section of Tertiary rocks of that region.

"Probably both Pleistocene and Pliocene—Basaltic lavas

——Marked erosional unconformity

Pliocene (?)-Andesitic lavas and agglomerates, with minor bedded material. (A little glassy rhyolite occurs in this group in the southern part of the Hawthorne quadrangle.)

——Erosional (in places angular) unconformity

Pliocene or upper Miocene—Rhyolites and quartz latites intrusive into and overlying the Esmeralda formation.

Upper Miocene (including beds considered to be of lower Pliocene age)—Bedded deposits and rhyolitic breccias with subordinate flows. This has been mapped in the field as 'Esmeralda formation.'

——Erosional unconformity

Miocene(?)—Lavas dominantly of andesitic composition but of considerable variety in places."

The most accurate correlation of the lavas southeast of Mono Lake is with those in the Hawthorne quadrangle lying adjacent on the north, for the late Tertiary basalt is continuous. The andesites and rhyolites lying beneath this basalt in the northern part of the area mapped by the author and in the southern part of the Hawthorne quadrangle mapped by Ferguson probably should be correlated. This correlation indicates that the lavas southeast of Mono Lake are younger than the uppermost Miocene sediments exposed in the Hawthorne quadrangle, mapped by Ferguson as "Esmeralda formation."

CORRELATION WITH SWEETWATER RANGE

Lying north of the town of Bridegport, California, and largely in the Bridgeport quadrangle (Fig. 1), the Sweetwater Range is about 35 miles west of Hawthorne, Nevada. W. H. Swayne (personal communication) has determined the following sequence of Tertiary rocks in that range; though the lacustrine series is lacking, correlation with Ferguson's section to the east seems good.

——Basin-Range faulting and consequent erosion

Basalt flows

——Erosional unconformity (?)

Andesitic flows, breccia, and tuff. These seem to be the same as those on the crest of the Sierra Nevada at Sonora Pass to the west since they are almost continuously traceable between the two areas.

——Erosional unconformity (?)

Rhyolite flows and tuff at least 200 feet thick

——Faulting, mineralization, and erosion

Older Tertiary lavas. These are chiefly andesite flows and breccias with a few shallow rhyolitic intrusions and one recognized rhyolite flow. All are hydrothermally altered.

——Very marked angular unconformity

Pre-Tertiary rocks

TABLE 2.—*Suggested Correlation of Tertiary Rocks*

	Southeast of Mono Lake	Hawthorne and Tonopah quadrangles	Sweetwater Range	Sonora Pass
PLEISTOCENE	Rhyolite flows, domes, and ash Andesite domes Olivine basalt flows Bishop tuff	Basalt flows		
	BEGINNING OF BASIN-RANGE FAULTING	BEGINNING OF BASIN-RANGE FAULTING	BEGINNING OF BASIN-RANGE FAULTING	BEGINNING OF BASIN-RANGE FAULTING
PLIOCENE	Rhyolite domes, flows, and tuff of Glass Mountain and vicinity Olivine basalt flows	Basalt flows	Basalt flows	
	DISCONFORMITY	DISCONFORMITY	DISCONFORMITY	
	Andesite flows, breccias, tuffs, and conglomerates	Andesite flows, breccias, and tuff. Some rhyolite in southern part of the Hawthorne Quadrangle	Andesite flows, breccias, and tuff	Andesite flows, breccias, and tuff
	DISCONFORMITY	UNCONFORMITY	DISCONFORMITY	
	Rhyolite tuff and flows	Rhyolite and quartz latite, extrusive and intrusive	Rhyolite tuff and flows	
MIOCENE		Continental sediments, upper part Esmeralda formation		
		DISCONFORMITY	UNCONFORMITY	
		Altered volcanic rocks, mainly andesites	Andesite flows and breccias with some rhyolite, altered and mineralized	
	MARKED UNCONFORMITY	MARKED UNCONFORMITY	MARKED UNCONFORMITY	MARKED UNCONFORMITY
	Pre-Tertiary rocks	Pre-Tertiary rocks	Pre-Tertiary rocks	Pre-Tertiary rocks

Correlation of the later Tertiary basalts, andesites, and rhyolites in the Sweetwater Range with similar rocks in the Hawthorne quadrangle suggests that their age is post-uppermost Miocene. This tentative correlation relates the late Tertiary andesites of this part of the Great Basin with those mantling much of the Sierra Nevada summit near Sonora Pass (Fig. 1).

AGE OF THE ESMERALDA FORMATION

First described by Turner (1900a, p. 197-208; 1900b, p. 168-170), the type section of the Esmeralda formation is in Big Smoky Valley at the north end of the Silver Peak Range (Fig. 1). The series consists of lacustrine and fluviatile sediments with intercalated volcanic rocks in its upper part. Apparently extending around the north end of the range, the formation is well exposed in the north end of Fish Lake Valley. Buwalda (1914, p. 335-363) traced these beds almost continuously from the type locality into Stewart and Ione Valleys near Cedar Mountain, 35 miles north (Fig. 1), and a vertebrate fauna collected from them there was determined as of upper Miocene age by Merriam (1916, p. 169-170). Since then, vertebrate faunas collected from the Cedar Mountain beds have been identified as both middle Miocene and lower Pliocene by Stirton (1932, p. 60-61); reexamination by Stirton (1939, p. 627) revealed that the middle Miocene fauna has been redeposited in this series of beds so that they must now be regarded as entirely lower Pliocene. Another vertebrate fauna characteristic of the whole Esmeralda section (2000 feet thick) in Fish Lake Valley has been identified by Stirton (1932, p. 60-61) as lower Pliocene. Axelrod (1940, p. 168) states that the flora from the Esmeralda type locality studied by Knowlton (1900, p. 209-220) with additional species more recently collected there indicate a lower Pliocene age.

Buwalda (1914, p. 351) extended his correlation of the Esmeralda formation to include beds of similar character and relationships near Hawthorne, 40 miles west of Cedar Mountain (Fig. 1). This was based on the similarity of the most abundant molluscan species collected near the Silver Peak Range, at Cedar Mountain, and near Hawthorne. Berry (1927, p. 1-15) has identified a flora from lacustrine beds 15 miles west of Hawthorne, in the Coal Valley locality. He states that this flora is "most certainly upper Miocene" and correlates it with the type Esmeralda since many of the species are identical. Axelrod (1940, p. 168-170) agrees with Berry by calling the Coal Valley flora "uppermost Miocene," but he states that the flora from the type section of the Esmeralda formation, examined by Knowlton, is distinct and younger. From the beds in Coal Valley, Stirton (1939, p. 634-635) obtained a meagre vertebrate fauna

probably representing the early lower Pliocene and which he thinks is older than the faunas from the Esmeralda formation near Cedar Mountain and in Fish Lake Valley.

AGE OF THE VOLCANIC ROCKS

Since at least part of the Coal Valley beds west of Hawthorne are early lower Pliocene or uppermost Miocene, the rhyolite, andesite, and basalt overlying them are probably lower or early middle Pliocene, though Ferguson suggests that the rhyolite may be upper Miocene. To the east near Cedar Mountain the similar andesites and basalts overlying the Esmeralda formation are certainly not older than lower Pliocene, based on vertebrate faunas. The lavas of these two areas are probably very nearly, if not exactly, the same age. The author logically assumes these rhyolites, andesites, and basalts were erupted contemporaneously with the older rhyolites, andesites, and basalts southeast of Mono Lake during the lower and early middle Pliocene. The rhyolite of Glass Mountain, younger than the Tertiary basalt flows, was erupted shortly after the basalt and is tentatively regarded as middle Pliocene.

STRUCTURE

GENERAL DESCRIPTION

Only the last event in a complex structural history—the faulting which has defined the present ranges—is here considered. Clearly it did not begin before the middle or late part of the Pliocene but continued until the present time.

Relative elevation of the ranges of the Great Basin along normal faults has now been clearly demonstrated by both stratigraphic and physiographic evidence. Southeast of Mono Lake, the displacement of Tertiary lava flows and tuffs affords conclusive evidence of the existence and location of the faults. These volcanic rocks, capping many of the ranges, occur in the adjacent downfaulted valleys, and other displacements can be clearly seen in many canyon walls. Where Tertiary rocks are lacking physiographic evidence must be relied upon. Those physiographic features affording the most conclusive evidence are: (1) the presence of a high, steep, little-dissected scarp with the lowest part of the adjacent valley at its very base; (2) the presence of a deeply weathered erosion surface on the summit of a relatively elevated range block; or a combination. The summits of some ranges which have no lava cappings are fairly smooth old erosion surfaces which could not form at the relatively elevated levels they now occupy. In contrast to the marginal scarps, these summit surfaces are more deeply weathered than the more

recently exposed scarps and represent parts of a once-continuous old age erosion surface now displaced by faulting.

East of Black Lake the scarp of the Benton Range is nearly 1000 feet high. Although battered by erosion, it is so little dissected that only incipient fans lie at its base and Black Lake stands very close to it. By comparison the eastern flank of the range, possessing a gentler slope is mantled in places by Tertiary lavas and tuffs tilted eastward and where these are absent is found an old erosion surface also inclined eastward. Blind Spring Hill is a small range relatively uplifted without tilting. The marginal scarps bounding it are slightly dissected while the summit is an erosion surface of moderate relief carved in pre-Tertiary granitic rocks and dikes intrusive into them.

The imposing scarps of the White Mountains and Sierra Nevada are higher (Pl. 2, fig. 1) as well as more rugged and deeply incised by youthful canyons which have dissected a once-extensive old erosion surface on the range summits. Because these high ranges intercept more precipitation, have larger watersheds, and were loci of Pleistocene ice accumulation, the greater dissection of their scarps resulted. At several places on the Sierra Nevada crest are small cappings of Tertiary basalt similar to that in the relatively depressed area to the northeast.

The total displacement on a fault, the recency of movement, the intensity of erosion, and the resistance of the displaced formations influence the conclusiveness of the physiographic evidence of faulting. Smaller displacement and greater dissection of the fault scarp and the old erosion surface make the evidence more obscure and often inconclusive.

The gorges, which afford drainage from the relatively downdropped open valleys, also indicate faulting. One cuts through the Benton Range, draining flood waters from Black Lake eastward to Benton; another drains Blind Spring Valley, in which Benton lies, through the north end of Blind Spring Hill to Owens Valley. A gorge also cuts eastward through the south end of Blind Spring Hill into Owens Valley. Still another tiny gorge drains an otherwise closed basin at the north base of Granite Mountain through a small fault ridge into Adobe Valley. Such youthful gorges, in contrast to the large open valleys, result from the tendency of streams to maintain an integrated drainage which is broken by the fault movements. Obviously not all youthful gorges draining more open terranes are evidence of faulting.

Lastly, lines of seeps and springs commonly indicate the location of a fault. The most notable example from this area is the line of springs trending northwestward through the valley area near Pellesier Ranch. Only alluvium is exposed there but the logical explanation of half a dozen springs in line is the existence of a fault.

Only the certain faults have been mapped. In the Benton Range south of Black Lake (Fig. 5) and to a lesser degree in the northern part of the range, where the evidence is clearest, even small faults can be mapped but so many exist only the largest of them can be shown. Such areas suggest a complex mosaic of fault blocks over the entire region. Like the ranges, the large valleys are faulted. Hills of basalt projecting above the alluvium in Adobe Valley indicate that faults, possibly extensions of some of those mapped in the surrounding hills, have displaced its bedrock floor. Also in the area north of Bishop the surface of the Pleistocene Bishop tuff is scarred by small fault scarps which demonstrate displacements in the valley block, and at least one fault traverses the valley near Pellesier Ranch.

Those faults which have suffered the greatest displacements lie at the bases of the range scarps where, instead of being bared by erosion, their traces are continually being buried by alluvium. Since the dips of these faults can rarely be determined, they cannot be proven normal. However, dip determinations may sometimes be made on faults traversing the range blocks. The fault surface of a small north-south fault crossing the canyon about 1 mile west of Benton is revealed in the canyon walls where rhyolite tuff has been dropped against granitic rocks (Pl. 3, figs. 2, 3). The fault dips 65 degrees west and is normal. Other exposed faults are normal, dipping between 60 and 75 degrees. Since determinable faults are all normal, by implication those less well exposed are taken as normal. No faults with relatively low angle of dip were observed, so that the faults in the structure sections (Fig. 4) are all dipping steeply.

The White Mountains (Pl. 2, fig. 1) have been relatively uplifted between 7000 and 8000 feet, and near Long Valley the Sierra Nevada has been relatively elevated between 3000 and 5000 feet. The smaller ranges between these are not comparable in height except the Glass Mountain range, facing the Sierra across Long Valley, which is bounded on the south by a fault scarp between 2500 and 3000 feet high (Pl. 2, fig. 2). The Benton Range stands between 500 and 1000 feet above the valleys at its western foot, and Blind Spring Hill has a similar relative elevation. In places the total displacement along these ranges occurs essentially along a single fault; elsewhere a similar total displacement is distributed among many smaller faults. Near Black Lake, for example, the unbroken scarp of the Benton Range (Pl. 3, fig. 1) is apparently the result of movement along a single fault, while to the north adjacent to Adobe Valley the range is raised to a similar relative height along many smaller faults.

FIGURE 1. LOOKING NORTHEAST AT THE FAULT SCARP
Along the northern White Mountains.

FIGURE 2. LOOKING WEST ALONG THE FAULT SCARP
North of Long Valley. Sierra Nevada in distance.

MAJOR FAULT SCARPS

FIGURE 1. LOOKING NORTH AT THE BENTON RANGE
Black Lake in middle distance at base of fault scarp.

FIGURE 2. TWO TILTED FAULT BLOCKS
On east flank of Benton Range. Granitic rocks capped by rhyolite tuff.

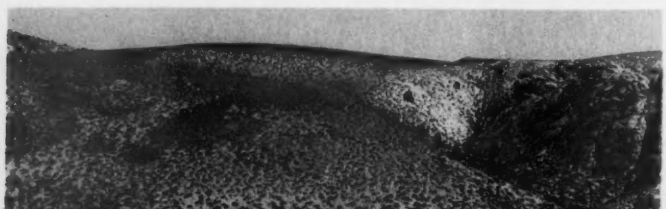

FIGURE 3. DETAIL OF FAULT SHOWN AT LEFT IN FIGURE 2.
Rhyolite tuff (center) dipping into granitic rocks (right).

FAULT MOSAIC

Throughout the whole region faults strike in nearly every direction, though commonly between northeast and northwest. These nearly straight faults intersect and cross in angular fashion making a fault mosaic. Intersecting faults are not noticeably offset one by the other, implying a negligible strike-slip component and a slip commonly down the intersection of the two faults. Such faults probably formed contemporaneously.

Though the individual faults are nearly straight, the lines of major displacement, shown by the fault scarps, bend in an angular or zigzag fashion. Such scarps are defined by more than one fault, the major displacement following first one and then another of different strike. Where an angular re-entrant occurs in a scarp one or both of the faults defining the sides of the re-entrant may continue with diminished displacement into the range (Fig. 5).

The western scarp of Black Mountain is defined by north- and northeast-trending faults so that the scarp alternates between a north-south and northeast trend. Less precise in pattern, the fault scarp forming the west side of the Benton Range is also irregular with angular salients and re-entrants (Pl. 3, fig. 1). The scarps along Blind Spring Hill and the small scarps which scar the surface of the Bishop Tuff to the south are likewise examples.

Thus, although the Benton Range, like many other ranges in the Great Basin, trends north-south for some 20 miles, few of the faults along which it was relatively elevated have that trend. The White Mountains trend north-south, yet the fault scarps along it are not straight because they are defined by faults some of which strike at distinct angles to the north-south trend. The same is true of the Sierra Nevada and the Walker Range to the north near Hawthorne. Clearly the stress condition which existed throughout this region at the time of faulting was relieved by normal displacements along zones trending, like the ranges produced, between north and north-northwest. The many faults not following this trend suggest some pre-existing structural element which controlled the location and trend of individual surface faults. Presumably the stress control of the north-south zones of major displacement lay at depth while the element which governed the observed individual faults existed nearer the surface. To what depth this surface control may have extended cannot be determined.

Hulin (1931, p. 307) has suggested the term "subsequent faults" for displacements controlled by "previously existing planes of weakness."

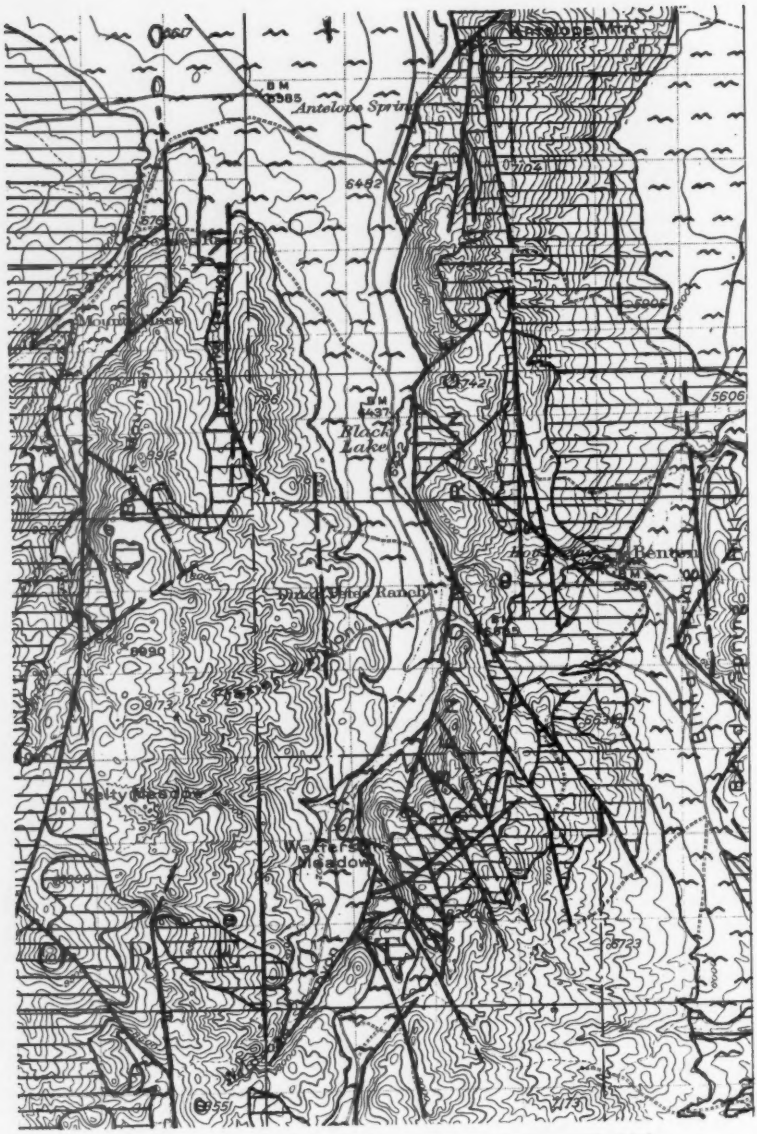

FIGURE 5.—*Topography and geology of Benton Range and vicinity*

Blank areas—pre-Tertiary rocks; lined areas—Tertiary volcanic rocks; wavy dashed lines—alluvium.

He took the Benton Range for the type region of "subsequent faulting" and suggested that the controlling planes of weakness might be defined by "joint planes, joint systems, or previously existing fissure systems." Both Mayo (1937, p. 182-183) and Chelikowsky (1940, p. 430-431) picture a similar control by joint systems along the Sierra Nevada fault scarp near Long Valley and control of these joints by still earlier structures. Spurr (1905, p. 79) suggested that zigzag fault scarps might be controlled by previously existing fractures, although he also suggested, without necessarily implying control by earlier structures, that they could be formed as lines of "equal dislocation" produced by intersection of two fault systems.

In the area under discussion these hypotheses may be based upon the following theoretical considerations:

(1) If any planes of relative weakness are oriented so that displacements along them will relieve the stresses affecting the region, whatever faults form will develop along them.

(2) If an essentially homogeneous structural system fractures, the fractures follow approximately straight lines or smooth curves. The lines of major displacements being angular indicate an element of heterogeneity in the structural system.

(3) Southeast of Mono Lake, the rocks displaced are primarily granitic with a thin capping of extrusive rocks. The effect of these lavas upon the faulting could only have been an unimportant very shallow control. Hence the controlling element must be heterogeneity within the plutonic and metamorphic rocks, several rather uniform intrusive masses and very small areas of metamorphic rocks. The irregular outlines of these intrusions do not control the faulting, for the fault pattern has no apparent relation to their contacts. In a structural sense, joints and earlier faults in the plutonic rocks are certainly a second feature of heterogeneity; probably these fractures defined planes of relative weakness followed by later normal faults, for, in a region literally shattered by joints and displaced along later normal faults, it seems reasonable that the faults should utilize earlier fractures instead of ignoring them.

(4) If they controlled the faults the joints necessarily existed prior to the displacements. They are older than the faults in this region, for the weathering of the old erosion surface, in places covered by Tertiary lavas and displaced by the faults, has gone deep along these joints isolating many large residual boulders of granitic rock.

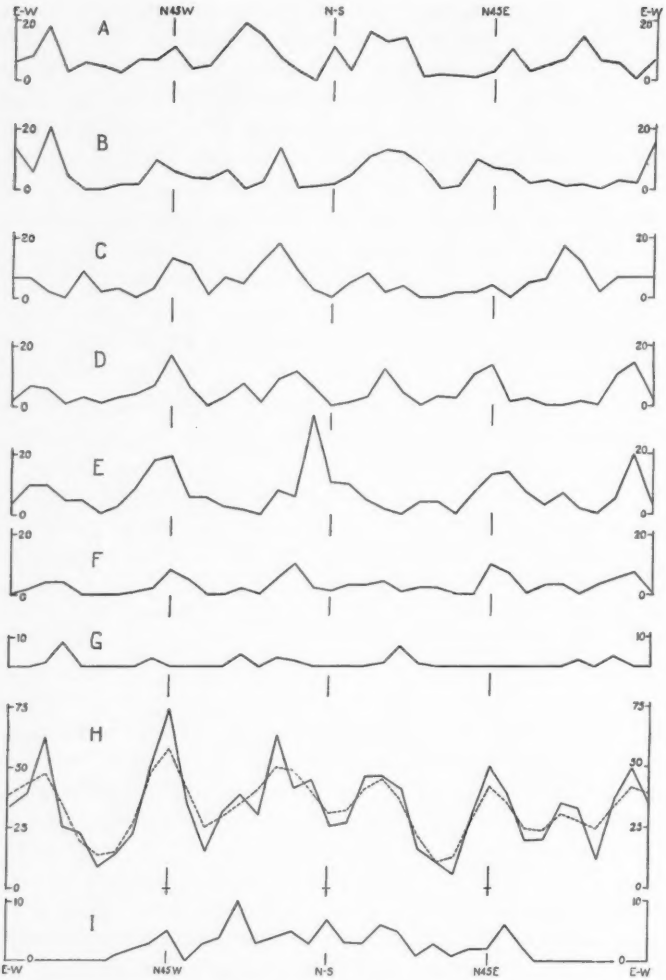

FIGURE 6.—*Strike of jointing in pre-Tertiary rocks of Benton Range and vicinity*

Abscissa—strike; ordinate—number of joints. Curve A represents joints east of Black Lake, B those west and southwest of Benton, C those near basalt southwest of Benton, D those east and northeast of Clover Patch, E those several miles east and southeast of Clover Patch, F those between Black Lake and Black Mountain, and G those on Blind Spring Hill. Curves A to E represent areas progressively from north to south. Curve H is a composite of A—G, and I represents the trends of faults in the area.

JOINTING

In an effort to determine whether any clear relationship exists between the joints and later faults, the strikes of many steep joints were determined. A clear understanding of the results involves a knowledge of the method employed in recording the joints in the field:

(1) Only joints with a moderate or high angle of dip were considered and only those well enough developed and exposed to afford a good determination of strike and dip were recorded.

(2) All steep joints could not be recorded, and therefore some selection was necessary. Separate groups of observations were made by traverses across unit areas distributed rather uniformly from north to south through the Benton Range. Additional readings in other parts of the region were made to determine the presence of any uniform joint system. In each unit area the author attempted to record all the well-exposed joints traversed and also to evaluate in a purely qualitative manner the relative abundance and continuity of joints in each set.

(3) From each group of observations a tabulation of the number of joints having each strike was made (strikes were determined to the nearest 5 degrees). This tabulation, shown for each unit area by curves A—G (Fig. 6), was combined into one composite curve (H). For each unit area the number of joints with a certain strike was translated into a percentage of the total number recorded there. Thus each unit area receives equal weight in a composite curve although the number of readings in the various areas is not identical; the curve based on percentage is essentially the same as that based on number of observations (H) and therefore is not reproduced.

(4) The dashed line curve (H) represents the data after smoothing. This smoothing was performed in the customary manner, according to the formula:

$$x = \frac{2y + y' + y''}{4}$$

where x is the new value for any point y on the solid-line curve and y' and y'' are the points adjacent to y. This is justified because there is a possible error of nearly 5 degrees in the strike of a joint if its true strike lies between two recorded directions and because since all joints could not be observed the data are incomplete.

These curves (Fig. 6) reveal the following facts:

(1) In each unit area the joints trend in nearly every direction.

(2) In each area certain trends are followed by a greater than average proportion of joints. These most prominent joint sets are not precisely defined but range in trend through angles of 10 or 15 degrees.

(3) Between areas represented by curves A—G there is some variation in the trends and prominence of corresponding joint sets. The most consistent is the set trending northwest. The north-northwest set is consistently present but becomes progressively more nearly north-south in trend as one proceeds from north to south (A to E). A set trending between N 10° and 20° E is represented only in the Benton Range north of Clover Patch (curves A, B, C, D) and seems to become more nearly north-south to the south. Joint sets striking northeast are not consistent in trend or prominence from one diagram to the next, while those trending nearly east-west are present everywhere.

(4) The composite diagram suggests a regional systematic arrangement of joints composed of at least four and possibly six sets.

In addition, the joints in the granitic rocks west of Adobe Valley appear to have a similar systematic arrangement. Diagrams representing the joint strikes in the Sierra Nevada near Long Valley and published by Mayo (1937, p. 180) are also similar, and Chelikowsky (1940, p. 432-433) states that the regional fracture system there has northwest, northeast, north-south, and west-northwest elements.

RELATION BETWEEN FAULTS AND JOINTS

The simplest visual comparison between fault and joint trends can be made by means of two curves, one (H) the composite curve representing all joint observations in and near the Benton Range, and the other (I) compiled from the strikes of the faults in the same region with the addition of faults from the northern and southern parts of the Benton Range beyond the area of joint observations. These additional faults increase the number recorded on the curve. The increase is justified because there seems to be a more or less constant regional joint pattern. Nevertheless, this fault curve may not represent all the faults in the area since it is based on few readings and since equal weight is not given to the various parts of the region, faults being more easily mapped in some areas than in others. The correlation of these curves can be summarized as follows:

(1) No faults approximate an east-west trend, supposedly because fault displacements with that trend could not relieve the stress condition. There was an extension of the region in an east-west direction, and the normal faults which accomplished it broke along planes trending more nearly north-south.

(2) Within the northwest-northeast quadrant, the joint and fault curves seem to show partial correlation.

(3) In order to test this correlation statistically, the correlation ratio between the two curves, H (solid line) and I, was determined, and a value of 0.61 obtained. A correlation ratio of 1.0 indicates perfect correlation whereas a value of 0.0 indicates an utterly random arrangement. According to Brunt (1931, p. 157), "when R (the correlation coefficient) is small, there is only very slight correlation between the two variable characters considered, and it seems doubtful whether any serious meaning can be attached to values of R which are less than 0.5." The correlation ratio and coefficient are essentially alike, so that Brunt's statement implies that some significance may be attached to the value 0.61. Since no faults trending nearly east-west exist in the region represented by curves I and H, and since it is logical to assume that faults of such trend could not be expected, the correlation ratio between the central parts of the two curves was calculated, between N 55°W and N 55°E where faults are recorded. The value of the correlation ratio between the recorded fault and corresponding joint strikes is 0.70, thus suggesting at least partial control of the fault trends by the jointing. Some other controlling factor or factors may also have been in operation.

The faults used in drawing curve I include minor faults of no great apparent extent and with displacements as small as 100 feet as well as major faults of greater extent and larger displacement. It is not clear how these faults should be weighted. Most of the small faults could not be satisfactorily mapped, and therefore curve I includes a greater proportion of large faults than actually exists. One might logically expect a better correlation between joints and smaller faults than between joints and larger faults, and thus the observed correlation is the more notable.

RELATION BETWEEN RANGES

The White Mountains and Sierra Nevada, standing like buttresses on either side of this region, seem to control the trend and location of the smaller ranges between them. Blind Spring Hill parallels that portion of the White Mountain scarp adjacent to it. The Glass Mountain range which is the only major one with an east-west trend stands facing that part of the Sierra Nevada scarp which trends east-west, and eastward where the Sierra Nevada scarp swings southward the faults along the Glass Mountain range do so likewise. The northern end of the Benton Range swings markedly to the east as though controlled by the eastward swing at the northern end of the White Mountains. Perhaps this suggested control of the minor ranges by the larger ones is more apparent than real.

SUMMARY OF CONCLUSIONS

Southeast of Mono Lake the Tertiary volcanic rocks, probably lower and middle Pliocene in age, are earlier than the normal faulting which produced the existing ranges. Some evidence supports the hypothesis that individual faults, along some of which these ranges were relatively elevated, were controlled by joints in the older rocks.

WORKS TO WHICH REFERENCE IS MADE

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BOXELDER BUTTE FROM THE NORTHWEST

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STRUCTURE AND PETROLOGY OF THE BOXELDER LACCOLITH, BEARPAW MOUNTAINS, MONTANA

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ABSTRACT

A partly eroded laccolith of the Shonkin Sag type about 450 feet thick and 2 miles across forms a prominent flat-topped butte a few miles east of Boxelder, Montana. An upper layer of cliff-forming syenite about 120 feet thick grades downward to a coarse, friable shonkinite about 175 feet thick. Below this are a cliff-forming shonkinite about 40 feet thick and a friable, ribboned shonkinite about 50 feet thick, which has horizontal layers of syenite. The layers (ribbons) of syenite are thin and closely spaced near the base of the laccolith and become thicker and more widely spaced upward. Narrow irregular dikes of aplitic syenite cut the syenite and shonkinite. The coarse, friable shonkinite contains syenitic segregations.

Qualitatively, the rocks are much alike mineralogically. The shonkinite contains augite, olivine, biotite, sanidine, plagioclase, and analcime. The syenites lack olivine and contain aegirite. The aplitic syenite lacks plagioclase.

The main part of the laccolith is believed to have been injected as a shonkinitic magma and to have differentiated in place by crystal settling to form the upper syenite and most of the underlying shonkinite. The ribbons of syenite and the dikes of aplitic syenite were formed by local segregations of the residual liquid from the crystallizing shonkinite. When the first magma had largely crystallized, a second injection of magma formed the brim shonkinite.

INTRODUCTION

LOCATION AND GENERAL STATEMENT

The Boxelder laccolith is in north-central Montana on the western fringe of the Bearpaw Mountains, 3 miles east of Boxelder, on State Highway Number 29 (Fig. 1). It is a prominent flat-topped though eroded butte locally known as Square Butte, or Boxelder Butte, and is the "bear's paw" for which the adjacent mountains are named.

The Bearpaw Mountains occupy about 800 square miles, and the western peaks rise to elevations of 6000 to 7000 feet. Andesitic, latitic, and phonolitic volcanic rocks and granular intrusive rocks with alkalic affinity make up more than half the mountainous area, and deformed sedimentary rocks of upper Cretaceous and Paleocene ages constitute the remainder. Dikes and sills are common in the plains area immediately adjacent to the mountains on the west and south. A brief reconnaissance report on part of the igneous rocks of the mountains was published by Weed and Pirsson (1896), and a brief summary account of the general petrology prepared by the writer was incorporated in a paper by Larsen (1940). A more detailed description of the petrology is intended for a forthcoming paper.

FIELD WORK AND ACKNOWLEDGMENTS

In the summer of 1938, before studying the Boxelder laccolith, the writer spent 4 days in the Highwood Mountains with C. S. Hurlbut, Jr., N. L. Bowen, J. F. Schairer, and D. T. Griggs in order to review the problems of the laccoliths of that area. During the same summer, Boxelder Butte was mapped topographically by a party of undergraduates from Harvard University; and for their work acknowledgment is given to E. L. Cherbonnier, Bernard Fisher, J. B. Lyons, T. C. Marvin, and R. G. Worman.

The writer spent about 10 days studying the geology of the laccolith. The topographic party spent about 2 weeks making the map reproduced in Plate 6. Control on the butte was by telescopic alidade and plane table, and the elevation of the initial base line on the West Ledge was determined with a barometer calibrated upon a bench mark in the town of Boxelder.

Financial aid from the Department of Mineralogy and Petrography of Harvard University made the study possible. Professor E. S. Larsen gave valuable assistance during the preparation of the manuscript, and, together with Professor C. S. Hurlbut, Jr., and David Griggs, offered helpful suggestions throughout the work. Mr. Edward Schmitz prepared the line drawings.

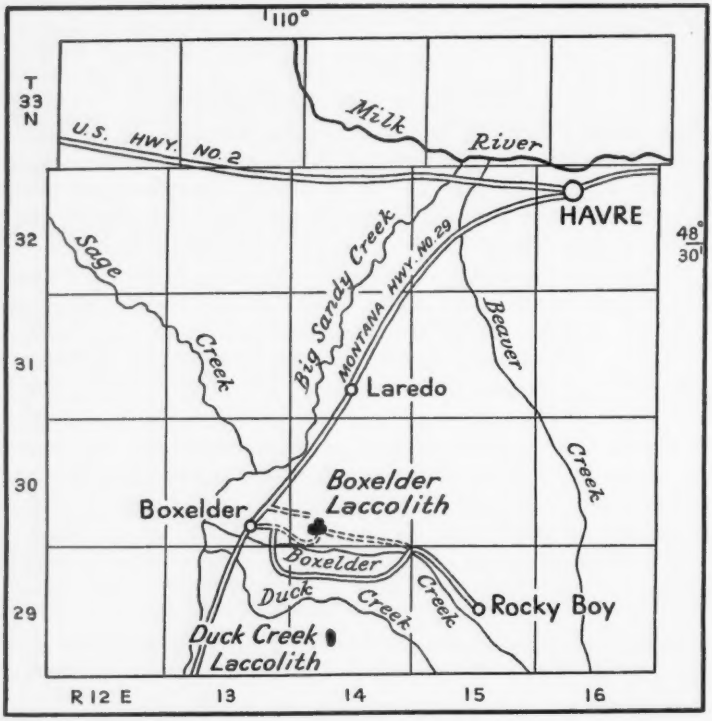

FIGURE 1.—*Index map of part of north-central Montana*
Showing general location and accessibility of Boxelder laccolith.

TOPOGRAPHY OF BOXELDER BUTTE

The butte, marked by precipitous cliffs on the northern and western sides, is visible from the plains for a great distance. Its summit stands at about 3600 feet, or 800 feet above the general level of the plains to the west. The upper 300 to 400 feet of the northwestern side of the butte is composed of granular igneous rocks, which because of a well-defined

FIGURE 1. NORTHWEST ESCARPMENT OF BOXELDER BUTTE IN BACKGROUND
Showing relation of topography and layered members of laccolith. Spheroidally weathered rock at top of lower cliff represents brim shonkinite. (*See* Figure 3.)

FIGURE 2. FACE OF SOUTH CLIFF OF WEST LEDGE
Showing multiple intrusions (dark-colored) and bedded sandstone of west border of laccolith.

NORTHWEST ESCARPMENT AND WEST LEDGE

FIGURE 1. THIN AND CLOSELY SPACED RIBBONS
Near basal contact of small shonkinite sill.

FIGURE 2. RIBBON NEAR TOP OF RIBBONED SHONKINITE
About 45 feet above base of laccolith.

RIBBONS (LAYERS) OF SYENITE IN SHONKINITE

layering and differential resistance, give rise to three cliffs: (1) an upper cliff of syenite (50 to 75 feet), (2) a lower cliff of medium-grained shonkinite (75 to 100 feet), and (3) an intermediate, less precipitous cliff of coarse-grained crumbly shonkinite (150 to 200 feet). Photographs of the three cliffs are shown in Plates 1 and 2. The intermediate cliff grades off toward flat ledges of harder shonkinite to the west and north. On the southern side of the butte the base of the upper cliff grades southward on coarse shonkinite and merges with the extensive South Bench, underlain by sandstone.

SEDIMENTARY ROCKS

The igneous rocks of the Boxelder laccolith have been injected into thick-bedded sandstone near the top of the upper Cretaceous Judith River formation, the general lithology of which in this region has been described by Reeves (1924). The sedimentary roof has been eroded except near the eastern and southern edges. On the eastern edge, the beds concordantly overlying the igneous rock dip eastward at angles up to 30 degrees and flatten out within half a mile. If projected westward up the dip, these beds would clear the topmost exposure of syenite on the summit of the butte by several hundred feet. It is reasonable to suppose, then, that either the beds originally flattened out over the top of the laccolith, as at Shonkin Sag, or, less likely, that a fault once was present in the roof west of these tilted beds.

On the South Bench, the sedimentary beds have an average dip of about 10° S., and, if projected northward, would clear the summit by a few hundred feet. The sedimentary beds on the northern slopes of the butte are relatively flat. On the western side the beds beneath the igneous rock dip a few degrees southeastward.

The border of the laccolith is exposed on the north and south sides of the West Ledge and on the gulley walls of the ledges on the north side of the butte. Structurally, these places are characterized by simple and multiple subordinate sills.

Contact metamorphic effects on sedimentary rock are negligible and, except for a noticeable hardening, jointing, and discoloration near contacts, the sandstone is practically unaltered. Beds composed of shells are very common, and plant fossils are plentiful in thin-bedded sandstone.

GEOLOGY OF THE LACCOLITH

FORM AND SIZE

In the Shonkin Sag laccolith, here considered the type for disk-shaped, layered laccoliths, the roof was uniformly raised, and arching was confined to the border of the intrusion, thus producing a flat-topped body

with steep slopes. Accommodation was further aided by thrusting at the border and by injection of minor sills beyond the border.

The Boxelder laccolith is larger than Shonkin Sag, is less well exposed, and is petrologically more complex. Some of its structural features are common to those in Highwood laccoliths. The border of the laccolith is

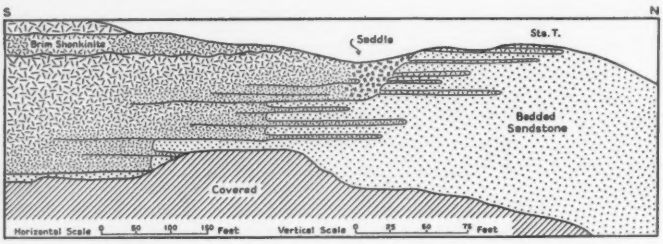

FIGURE 2.—*Border of Boxelder laccolith near triangulation station* T

Showing schematically fringing sills, overriding sill-like bodies, vertical abutments, and fractured (double track pattern) shonkinite. Hachures represent chill facies and contacts between sills.

characterized by multiple sills (Pl. 4; Fig. 2). The topmost sills are covered by beds of sandstone stratigraphically equivalent to the roof strata of the laccolith. On the walls of the gulleys on the south the thickest sill is about 50 feet.

A reconstructed, in part idealized, cross section of the laccolith is shown in Figure 13, and a geologic map is shown in Plate 6.

The area now underlain by igneous rock, excluding fringing sills, is about 1½ square miles, or less than half the original laccolith. The thickest igneous section is about 385 feet, and the original thickness was possibly 450 feet. The true center of the laccolith is not exposed, but the upper syenite cliff between stations L and N is presumably a few hundred feet away from the center.

EXPOSED VERTICAL SECTION

A generalized section of the laccolith showing the layering is indicated in Figure 3. The summit and upper cliff or about one-fourth the exposed column are composed of syenite. The remaining three-fourths is shonkinite whose layers have been named, in order below the syenite: friable, brim, ribboned, and rubbled shonkinites.

Six vertical sections are represented in Table 1. These are located with respect to the triangulation stations. Any measurement of the thickness of the friable shonkinite must be carried over a horizontal distance of at least 500 feet.

THICKNESS IN FEET	SECTION	NAME	REMARKS AND DESCRIPTION
ROOF		SANDSTONE	Sandstone and shell beds
50-100		UPPER RIBBONED SHONKINITE	Presence interpreted Increase of grain size of rock downward from roof of laccolith;
100-120		SYENITE AND MAFIC SYENITE DIKES	Increase downward in grain size and content of dark minerals. Local patches of pegmatitic syenite. Few horizontal dikes of aplitic syenite free of plagioclase;
150-200		FRIABLE SHONKINITE	Decrease downward in grain size and increase downward in content of dark minerals. Local patches of segregated syenite. Small discontinuous dikelets of syenite free of plagioclase;
40-60		BRIM SHONKINITE	Younger intrusions of shonkinite. Rock in contact with friable shonkinite near center of laccolith has highest specific gravity
30-50		LOWER RIBBONED SHONKINITE	Increase in grain size upward from base. Many well-defined layers (ribbons) of syenite which increase in spacing and thickness upward from base and inward from border of laccolith. Ribbons are irregular and analcime-rich near top;
10-30		RUBBLED SHONKINITE	Locally fractured or ribboned.
FLOOR		SANDSTONE	Sandstone and shell beds.

FIGURE 3.—*Columnar section of igneous rocks of Boxelder laccolith*

Two horizons are well exposed. The first, and more pronounced, is the contact of the syenite and underlying shonkinite, everywhere represented by a break in slope. The second is the hard brim shonkinite represented

by the top of the lower cliff. The upper horizon was used as a datum plane of reference for all the vertical sections of the laccolith and for the specific gravity and heavy mineral curves discussed later in this paper.

RIBBONED SHONKINITE

The most characteristic feature of the shonkinite at the base and border of the laccolith, in the many fringing sills, and in many dikes and sills

TABLE 1.—*Some measured sections of igneous rocks, Boxelder laccolith*
(Approximate thicknesses in feet)

		West of Station G	West of Station H	West of Station N	West of Station L	East of Station P	East of Station A
To summit of butte from top of upper cliff	Syenite	40	45	40	35	40	?
Upper cliff	Syenite	55	65	50	70	45	?
Intermediate cliff	Friable shonkinite	165	200	165	175	150	?
Lower cliff	Brim shonkinite	60	45	50	50	50+	65
	Horizontal injection series of ribboned and rubbled shonkinite	40	30	60	?	50+	40
Total thickness of sections in feet		380	385	365	330+	335+	?

in the adjacent plains area is the presence and arrangement of light-colored ribbons on the surface. The ribbons are thinnest, most closely spaced, and more generally parallel near chilled margins (Pl. 3).

The ribbons are exposed edges of thin layers of syenite in shonkinite. In large dikes, and in the top part of the ribboned shonkinite, the ribbons are irregular and interconnected. The widest ribbons are about 3 inches, and one is shown in Plate 3. A detailed record of ribboned shonkinite in two places at the lower part of the laccolith follows:

LOCALITY 1, RIBBON COUNT IN SILL B OF FIGURE 7

h. 10 feet, few ribbons, 1 to 2 inches thick and 1 to 2 feet apart.
g. 2 feet, few ribbons, ½ to 1 inch thick and 3 to 5 inches apart.
f. 5 inches, one ribbon, ½ inch thick.
e. 6 inches, 3 ribbons, ¼ inch thick and 2 inches apart.
d. 6 inches, 7 ribbons, ⅛ inch thick and 1 inch apart.
c. 6 inches, 12 ribbons, 1/32 inch thick.
b. 6 inches, few ribbons, white lines in fine-grained rock.
a. Irregular base of sill, chill facies of shonkinite.

Ribbons make up about 7 per cent by volume of measured vertical section of shonkinite at this locality.

LOCALITY 2, RIBBON COUNT IN SILL C OF FIGURE 8

g. base of the hard "brim shonkinite."
f. 10 feet, irregular and interconnected ribbons, up to 3 inches thick.
e. 15 feet, 4 wavy ribbons, 2 to 3 inches thick.
d. 10 feet, 5 ribbons, 2 inches thick.
c. 10 feet, 10 ribbons, 1 inch thick.
b. 1 foot, many ribbons, thin lines in fine-grained rock.
a. Irregular base of sill, chill facies of shonkinite.

Ribbons make up about 6 per cent by volume of measured vertical section of shonkinite at this locality.

RUBBLED SHONKINITE

Several of the sill-like bodies of ribboned shonkinite at the base and border of the laccolith are more or less well rubbled. Younger bodies of ribboned shonkinite intrude or directly overlie such bodies of rubbled shonkinite and locally are also fractured. Some shonkinite sills are rubbled in some places and ribboned in others. Some fragments of the rubbled shonkinite themselves contain layers of syenite. It is evident that the layering (ribboning) took place before the wholesale fracturing in these sills. Not uncommonly, the spaces between fragments of rubbled shonkinite are filled with platy calcite and glassy analcime.

TONGUES OF RIBBONED SHONKINITE

In many exposures on the face of the lower cliff, small and irregular sill-like bodies of a stony or fine-grained ribboned shonkinite cut ribboned or rubbled shonkinite. A few types of these tongues are shown in Figure 4. In a few places they can be traced into larger bodies of ribboned shonkinite and are therefore interpreted as the chilled, frontal parts of migrating bodies of shonkinitic magma.

FRINGING SILLS OF RIBBONED SHONKINITE

Along the steep slopes of the spurs and ledges on the northern, southwestern, and eastern parts of the butte, well exposed simple and multiple sills of ribboned shonkinite are not uncommon. The simple sills are mostly less than 4 feet thick and are most numerous in the thick-bedded sandstone and shell breccia. Each sill maintains its thickness between beds and ends abruptly in a blunt nose. The very large sills, some of which are between 20 and 50 feet thick, extend for a few hundred feet beyond the main border of the laccolith.

In some places, two or more sills in contact with each other merge inward toward the main body of the laccolith to form a single multiple sill and farther inward to form a simple sill. Mutual contacts of multiple sills always disappear inward toward the center of the laccolith. The internal relations of one multiple sill on the face of the cliff east of station A are shown in Figure 5.

Some of the fringing sills which end in blunt noses pick up again within a few feet in a series of nodes. This feature was well described by Griggs (Hurlbut and Griggs, 1939) for the fringing sills of the Highwood laccoliths and was attributed by him to the irregular fingering of the front of a sill during its implacement.

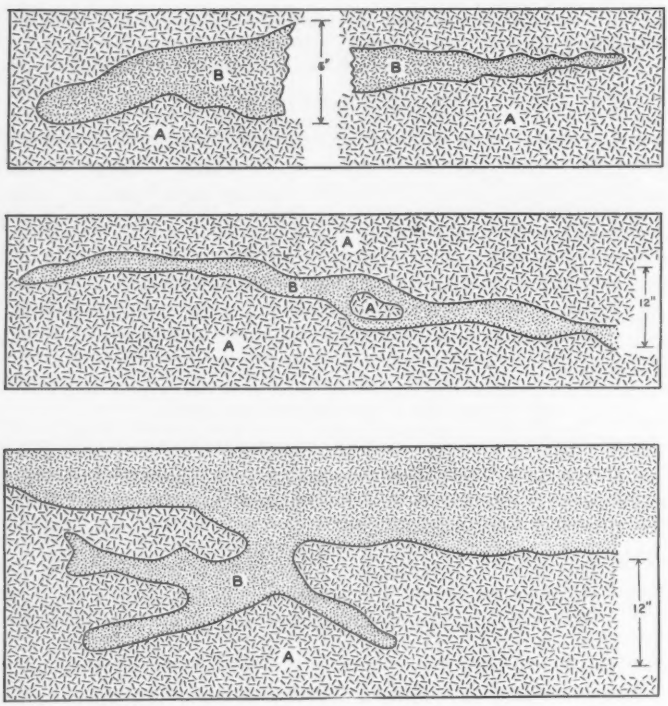

FIGURE 4.—*Tongues of fine-grained shonkinite* (B)
In less fine-grained and more altered ribboned shonkinite (A); in some places shonkinite of (B) likewise contains ribbons.

Where sills end in blunt noses, the ribbons are parallel to the curvature of the nose. Similar ribbons are also present within the nodes beyond the noses of sills. The sandstone in the immediate vicinity of the noses is fractured, but more intensively so between nodes. The bending of sedimentary beds between nodes, as described by Griggs and called "breccia ridges", was not observed in the Boxelder laccolith.

INCLUSIONS

Many rounded to subrounded inclusions of fine-grained shonkinite in ribboned shonkinite have been observed; in all cases the inclusions are

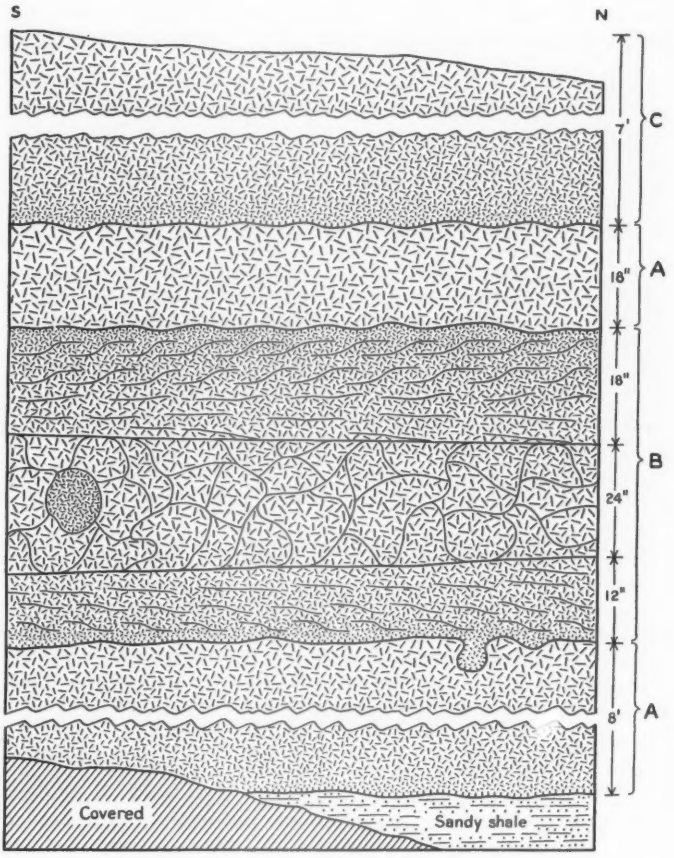

FIGURE 5.—*Multiple shonkinite sill*

Exposed on eastern wall of spur at triangulation station A. All three sills are ribboned. Sill (B) is divided into three segments, and ribbons in each are shown schematically by lines drawn within sill.

relatively fresh, and the host rock is altered. Inclusions range from 1 to 12 inches, but those between 3 and 5 inches are most common. The syenitic ribbons in the host rock, as shown in Figure 6, do not traverse

the included fragments but are deflected around them. Inclusions were not observed in the upper part of the friable shonkinite or in syenite.

HORIZONTAL INJECTION SERIES

Complex intrusive relations between successively younger sill-like bodies of ribboned shonkinite are well exposed at the base and border of

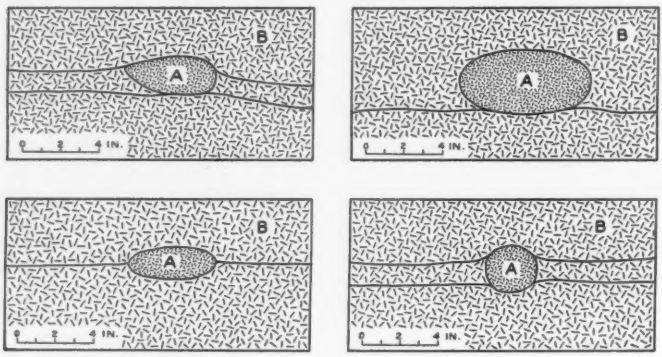

FIGURE 6.—*Inclusions of shonkinite* (A)
In coarse-grained ribboned shonkinite (B). Syenitic ribbons are deflected around borders of inclusions.

the laccolith. The relations at a few localities are illustrated by Figures 7, 8, and 9.

ILLUSTRATED SECTIONS OF CLIFF

The lower part and the border of the laccolith are composed of a complex series of horizontal intrusions of ribboned shonkinite. The lowest part comprises a group of rubbled shonkinite sills overlain irregularly by a group of younger ribboned shonkinite sills. Between the two groups lenses, wedges, or displaced blocks of fractured sandstone are common. The sills are progressively younger upward in the section. The exposed sills, however, are not complete; for, although each rests on an older one and is chilled at its lower contact, the top of no one sill is present on the cliff face. Grain size increases upward from the chill facies in each sill, and the thickness and spacing of the syenitic ribbons likewise increase upward. The question which naturally arises is: what has happened to the top part of each sill?

Many favorable exposures show that the contacts between sills disappear inward from the border toward the center of the laccolith. Along any one horizon the ribbons also become irregular and interconnected. Two relations control the arrangement of the ribbons: (1) The vertical

distance above the bottom contact of a sill, and (2) The horizontal distance inward from the border of laccolith.

The outer, frontal margins of large sills at the border of the laccolith are rubbled in places and enclose small tongues of ribboned shonkinite.

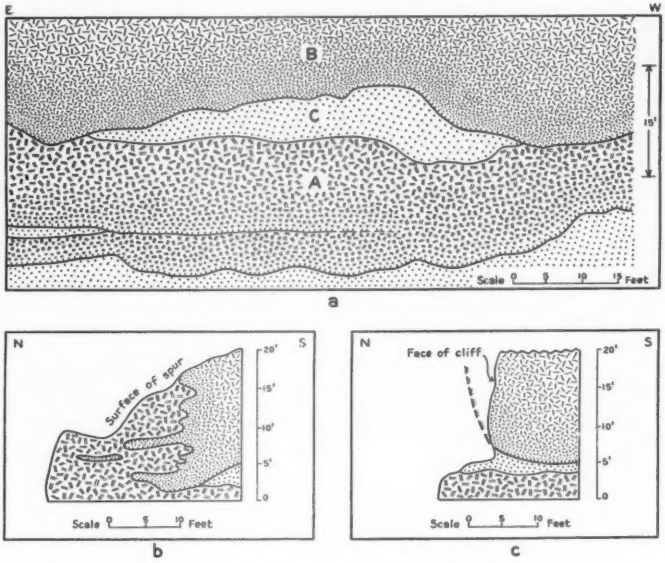

FIGURE 7.—*Lower cliff of shonkinite east of triangulation station* A

Single tracks represent ribboned shonkinite, and double tracks rubbled shonkinite. Sketch a represents cliff, and sketches b and c illustrate faces of two spurs. Stippled area (C) represents an irregular lens of fractured sandstone. Only lower part of each sill is present on cliff.

UPPER RIBBONED SHONKINITE

Three lines of evidence, all indirect, indicate that a layer of shonkinite once existed above the syenite: (1) From analogy with the Shonkin Sag laccolith, an upper shonkinite possibly measured between 50 and 100 feet. (2) The general grain size of lower shonkinite increases upward, so that there must have been a counterpart at the top of the laccolith. (3) The tops for the sills at the bottom of the laccolith must have been somewhere above the syenite and below the roof of the laccolith. Thus, the Boxelder laccolith, before erosion, was enveloped or encased by ribboned shonkinite (Fig. 13).

SYENITE AND MAFIC SYENITE

General statement.—The syenitic member of the laccolith comprises both the mafic and less mafic syenite and includes all the rock above the base of the upper cliff.

Distribution and thickness.—The summit of the butte is plagioclase syenite, although much of the bedrock is concealed under a thin veneer of till. The topographic break at the base of the syenite is persistent almost entirely around the butte. The greatest exposed thickness of the

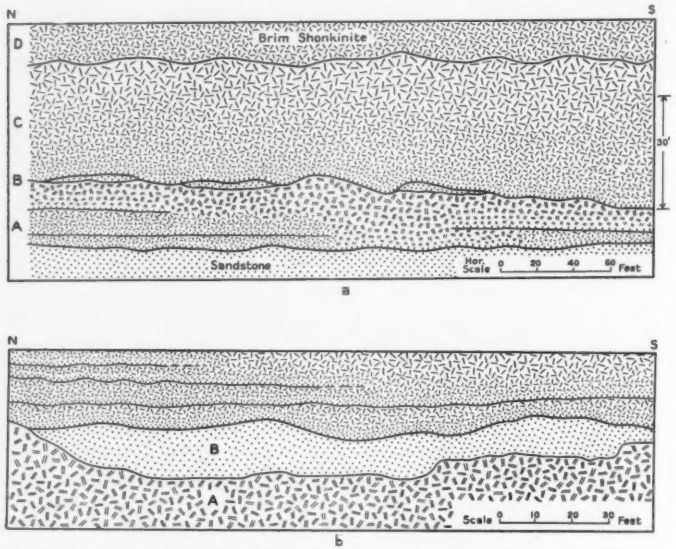

FIGURE 8.—*Lower cliff of shonkinite west of triangulation station* N

Sketch a shows four, well-marked horizons on face of cliff. Horizon A composed of multiple sills of ribboned shonkinite locally rubbled (double tracks). Horizon B represents irregular lenses of fractured sandstone. Horizon C represents a large multiple sill of ribboned shonkinite, and its lower part is shown in sketch b. Horizon D represents brim shonkinite. In each sill depicted above, only lower part is present on face of cliff.

syenitic member, located on the accessible western part of the butte, measures about 110 feet and is probably within 10 to 20 feet of the original thickness. The syenite thins eastward to about 10 feet over a distance of 1½ miles. The marked eastward slope of the summit of the butte may in large part be a stripped surface of a lens-shaped syenitic unit, for the syenite is much more resistant to erosion than the shonkinite below it and that which is interpreted to have been above it.

Contact of syenite with shonkinite.—The contact of syenite with underlying shonkinite is excellently exposed for more than 3 miles circumferentially around the base of the upper cliff and is everywhere marked by a discontinuity in slope and a perceptible change in color of the weathered rock. From a distance this would suggest that the change

between syenite and shonkinite might be abrupt. However, careful studies in the field and laboratory show that there is a gradual change in the proportion of dark minerals from the shonkinite upward through mafic syenite to syenite. Much of this change takes place in a distance of

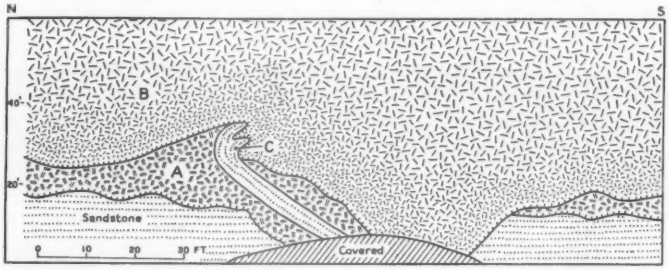

FIGURE 9.—*Face of lower cliff west of triangulation station* G

Relations of a large, sill-like body of ribboned shonkinite (B), locally multiple, are shown with a sill-like body of rubbled shonkinite (A) and a displaced block of fractured sandstone (C).

from 10 to 20 feet. This relation is true for more than a mile of exposure on the western side of the butte.

Near the edge of the laccolith the syenite and underlying shonkinite are less mafic than their equivalents on the western side of the butte. The contact, as shown in Figure 10, is not gradational. For a vertical distance of 3 to 15 feet numerous irregular and intricately connected stringers of analcime-rich syenite, satellitic but continuous from the syenite, penetrate downward into the shonkinite. The syenite here appears to have crystallized after the shonkinite.

At those places on the butte where the change of syenite to shonkinite is gradational, but not where the contact is a mixed zone of satellitic projections, there are at least three persistent, irregular, generally horizontal dikes of analcime-rich aplitic syenite in the mafic syenite.

Field description of the syenite.—The grain size of the syenite near the center of the laccolith (west cliff) is uniformly coarser than the syenite on the summit or on the eastern side of the butte. The syenite at the top of the butte is more compact than that on the face of the cliff, and the mafic syenite is more friable than either of them. The rock is chalky white on the weathered surface, largely due to the amount of zeolites present. The typical syenite has about 20 per cent dark minerals, chiefly biotite, aegirite, and aegirite-augite. Altered olivine is present in the mafic syenite but absent in the true syenite. Hornblende is rare. The syenite

above the shonkinite near the edge of the laccolith has a small amount of entirely altered olivine.

APLITE DIKES IN MAFIC SYENITE

A few irregular and essentially horizontal dikes of plagioclase-free aplite syenite are present in the mafic syenite on the face of the upper

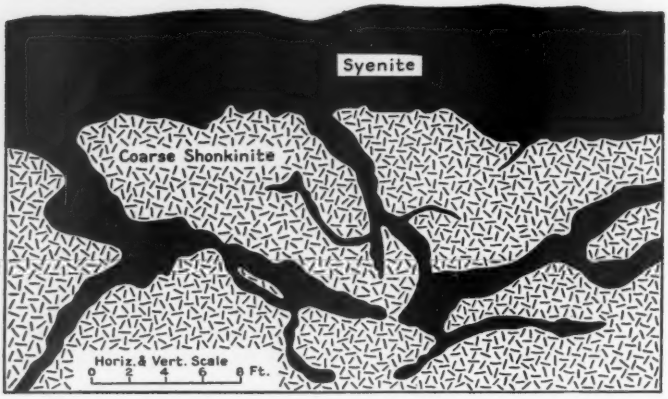

FIGURE 10.—*Irregular contact of syenite and underlying friable shonkinite*
North of triangulation station S.

cliff clockwise from stations Q to P. The largest of these dikes, about 3 feet thick, is exposed between stations L and F. No dike can be traced continuously for more than a few hundred feet, but all are wavy and interconnected by splits and stringers. One of these dikes is shown in Plate 4. The dikes are present only in the lower half of the syenitic member and only near the center of the laccolith. They thin out to a few inches near stations P and Q. In a measured vertical section of syenite west of station L, dikes make up about 5 per cent by volume of the vertical section of syenite.

For the most part the dikes are nearly horizontal but locally they bend down in troughs that are as much as 20 feet below the general horizontal plane. The lowest persistent dike does not itself transgress the contact of syenite with underlying shonkinite, but in a few places satellitic projections extend 50 feet into the top of the shonkinite. The contacts of the dikes with the mafic syenite are sharp. The interlocking texture of the sanidine in the dikes is more compact than that in the syenite. The rock of the dikes contains less femic minerals, more

aegirite, and more analcime than the true syenite; it contains no olivine or plagioclase.

One small dike of aegirite syenite in the Shonkin Sag laccolith was observed which, in its appearance in the field, resembles the aplite dikes in the Boxelder laccolith.

PEGMATITIC SYENITE

The syenite, and rarely the aplite dikes, have small scattered patches of pegmatite; but no persistent horizon of pegmatitic syenite exists in the Boxelder laccolith as in the Shonkin Sag laccolith. Generally, the pegmatites are most common on the upper cliff, between 20 and 50 feet below the very summit of the butte, and more than 40 feet above the syenite base.

The largest pegmatitic body found measured a few cubic feet. This body was localized at the contact of one aplite dike near the base of the syenite west of station G. The dike appeared younger than the pegmatite, but the relations were indefinite. Mineralogically, the pegmatite is similar to the host syenite but contains some hydrothermal minerals not found in the host. Sanidine, analcime, plagioclase, and biotite make up the bulk of the rock. The structure of the biotite is the most conspicuous feature of the pegmatite, and a photograph of a hand specimen is shown in Plate 4. Skeletal growths are randomly oriented and traverse large feldspar crystals.

Solution cavities are common in this and other pegmatites in syenite, and their walls are coated principally with analcime and other zeolites. An unidentified mineral occurring as radially fibrous bluish-white pellets attached to the walls of solution cavities is described in a later section on mineralogy.

FRIABLE SHONKINITE

This member everywhere forms the gentle slopes between the upper cliff of syenite and the lower cliff of brim shonkinite. The thickest section of friable shonkinite measured is about 200 feet, with the traverse carried along a map distance of about 500 feet. This member may be thicker at the center of the laccolith.

The friable shonkinite crumbles readily when struck with a hammer. Much of the intermediate slope of the butte has a thin mantle of loose sand caused by natural disintegration of the rock. Such a crumbly character is caused in part by chemical decay of salic minerals and by the lack of sufficient binder for the coarse, dark minerals.

The grain size of this shonkinite increases upward from the top of the lower cliff to the base of the upper cliff. Its lower part is in con-

tact with younger intrusions of the brim shonkinite and thus is separated from the top portion of the ribboned shonkinite. The latter in some places is also very crumbly and cannot be distinguished from the friable shonkinite. On some parts of the butte, ribboned shonkinite passes upward from the type with orderly ribbons to that with interconnected ribbons and finally passes into typical friable shonkinite with occasional dikelets. For purposes of description, however, the friable shonkinite is considered that portion lying above the hard shonkinite intrusions which hold up the brim of the lower cliff. The member is probably basin-shaped.

The rock of the upper part of the friable shonkinite member is more greenish than the main part due to the greater proportion of aegirite-augite with respect to the other femic minerals. Long and thin dikelets of aplitic syenite, apparently discontinuous, are abundant in the upper part, and some of them can be traced into the large aplite dikes in the base of the mafic syenite.

Two types of small syenitic bodies are present in the shonkinite. One is composed of irregular knots, shreds, and stringers of coarse-grained feldspathic rock which in every respect can be regarded as local segregation patches of syenite in shonkinite (Pl. 5). A second type is composed of small dikelets and pipes of coarse or aplitic syenite (Pl. 5). In general the coarse-grained bodies have gradational contacts, are only a few inches across, and contain plagioclase. The fine-grained bodies, although only a few inches across, are as much as several feet long, have sharp contacts with the host shonkinite, contain no plagioclase, and locally have a skeletal growth of biotite. Both types are rich in analcime, but the second type is commonly the richer. No estimate was made of the amount of individual syenitic bodies present in the shonkinite, but a few per cent is probably a maximum amount.

In some places the shonkinite has small patches richer in light minerals and, although these are spotty, they are not uncommon. In other places, small patches are extremely rich in dark minerals, and the rock is very crumbly.

On weathered surfaces of friable shonkinite a banded structure is visible, which is caused by the orientation of biotite. There is no completely systematic orientation in the butte, but at the contact of syenite and shonkinite the banding is generally horizontal, conforming to the major layering of the laccolith. No inclusions have been observed in the upper part of the shonkinite, but several shonkinitic inclusions were seen in the lower part.

FIGURE 1. LARGE APLITE SYENITE DIKE
In mafic syenite near center of laccolith.

FIGURE 2. SKELETAL GROWTH OF BIOTITE
In pegmatitic syenite near center of laccolith

APLITE DIKE AND SKELETAL BIOTITE

FIGURE 1. SEGREGATIONS OF COARSE-GRAINED SYENITE
In lower part of friable shonkinite.

FIGURE 2. SMALL VERTICAL SYENITE PIPE (FOREGROUND) IN FRIABLE SHONKINITE

One of several partly exposed bodies west of N.

SYENITE SEGREGATIONS AND SYENITE PIPE

BRIM SHONKINITE

Distribution and thickness.—The top of the lower cliff on the western side of the butte is made up of a hard shonkinite which separates the friable from the ribboned shonkinite. In some places (see Plate 2) the rock stands out boldly with its spheroidal weathering. The maximum thickness of this member of the laccolith is 60 feet. Because of the topographic position of the rock, this member is called the brim shonkinite. It is composed of more than one intrusion, but in the center of the laccolith the mutual contacts are not perceptible. Near the edge and beyond the border of the laccolith, the brim shonkinite is very fine-grained and is typically ribboned.

General description.—This extremely hard rock is rather homogeneous in appearance. The typical rock is medium-grained, heavy, and not so mafic as the friable shonkinite. Its contacts are not generally abrupt near the center of the laccolith. It cuts across the syenite ribbons in the underlying rock (east of station A). In some places segregation patches of syenite are present in the brim shonkinite. In a few vertical places there is a difference in color or grain size and this, in addition to petrographic data, indicates that the member is made up of more than one intrusion. West of station G, several interconnecting bodies of brim shonkinite are exposed in the friable shonkinite, and it is clear here that after the formation of ribboned and friable shonkinite a resurge of shonkinitic magma was injected at this horizon. Isolated, small bodies of brim shonkinite are common near the edge of the friable shonkinite.

PETROGRAPHY AND MINERALOGY

GENERAL PETROGRAPHY[1]

All the rocks of the Boxelder laccolith are partly altered, and chemical analyses are not warranted. In thin section, the mafic syenite and friable shonkinite are monotonously similar, consisting largely of pyroxene, olivine, biotite, sanidine, and plagioclase. The typical syenite differs from these rocks chiefly in having fewer dark minerals and in having more aegiritic pyroxene. Sanidine crystals dominate the texture, enclosing the other minerals poikilitically. The typical brim shonkinite differs from the friable shonkinite chiefly in having smaller grains of sanidine and plagioclase. The olivine in the brim shonkinite is less altered than in the other rocks.

[1] In future papers on the Bearpaw Mountains, plagioclase shonkinite will be petrographically distinguished from shonkinite, nepheline shonkinite, nepheline-sodalite shonkinite, and pseudoleucite shonkinite. This suggested classification is not in accord with Johannsen or Weed and Pirsson.

The chill facies of the ribboned shonkinite contains the same minerals as the other shonkinites, and all of the olivine and a part of the pyroxene appear to be intratelluric. The rest of the pyroxene grades seriately in size to fine crystals in the groundmass. Continuously away from the contact, the amount of coarse grains of pyroxene increases, but not their

TABLE 2.—*Approximate modal analyses and specific gravities of some rocks of the laccolith*

	1	2	3	4	5	6
Olivine	12	13	23	14	8	0
Pyroxene	25	20	18	35	20	6
Biotite	10	22	18	17	15	3
Plagioclase	n.d.	2	2	5	5	8
Sanidine and analcime	n.d.	n.d.	38	n.d.	49	83
Actual density	2.81	2.82	2.99	2.88	2.64	2.42
Recomputed density	2.88	n.d.	3.02	3.03	n.d.	n.d.

[1] Chilled ribboned shonkinite.
[2] Brim shonkinite, typical rock.
[3] Brim shonkinite, from top, near center of laccolith. Highest density of any rock in laccolith.
[4] Friable shonkinite, near bottom.
[5] Friable shonkinite, near top.
[6] Syenite, typical rock.

size. The biotite, sanidine, and plagioclase increase in grain size progressively away from the contact, the sanidine most rapidly.

The contacts of syenitic ribbons in shonkinite are well-defined in hand specimen and ill-defined in thin section. The ribbons actually are femic-poor streaks and contain no olivine or augite. Aegirite, aegirite-augite, and a very pleochroic biotite make up the small percentage of dark minerals present in the rock. The groundmass of sanidine and plagioclase seems to be continuous with that of the host rock, having the same grain size and texture. The largest ribbons are richest in analcime; some contain as much as 40 per cent, which is concentrated in the central part of the ribbons.

The rock of the aplite syenite dikes is similar to that of the ribbons except that it lacks plagioclase. The sanidine is highly altered in this rock, and analcime is abundant. The aplite syenite from the small bodies in the friable shonkinite has the same appearance in thin section as the syenite dikes.

Approximate modal analyses of some typical rocks with their measured and calculated specific gravities are listed in Table 2.

MINERALOGY

Thirty-nine thin sections of rocks from the Boxelder laccolith were studied in detail, with a distribution as shown below:

4	thin sections,	syenite and mafic syenite
13	" "	friable shonkinite
9	" "	brim shonkinite
8	" "	ribboned shonkinite
2	" "	syenitic ribbons
1	each of	aplite syenite dike in mafic syenite, dikelet in friable shonkinite, pegmatitic syenite.

From these sections and from powders of the rocks studied by the immersion method, many data were obtained on individual minerals and relations between minerals. The principal features brought out by these studies are:

(1) Grains of altered olivine are euhedral in the chill facies of shonkinite but are strongly resorbed within the main body of the laccolith. The olivine is probably intratelluric, for the size, shape, and kind are similar in all the shonkinites. For example, the olivine at the base of the friable shonkinite has a β refractive index of 1.705, and that in the brim shonkinite of 1.703, well within the limit of error of the method. The optical properties of this olivine indicate a ratio of Mg/Fe of about 78/22. The average length of olivine grains is between 1 and 2 millimeters, and the mineral makes up about 10 to 15 per cent of undifferentiated shonkinite.

The main alteration product of olivine is a reddish-brown, iddingsitic material with a high refractive index. This grades to a greenish, serpentine-like material with a minimum refractive index of about 1.530 and a moderate birefringence. Some of the alteration may be late magmatic, some hydrothermal, and some may be caused by surface waters. Aureoles of green and brown biotite are common around resorbed olivine. Coronas of biotite completely enclose many elliptical grains of resorbed olivine which are veined by the iddingsitic alteration.

(2) In the chill facies of shonkinite, less than 50 per cent of the pyroxene is in large euhedral crystals up to a few millimeters long. The pyroxene varies in composition from augite to aegirite. The largest augite crystals are at the lower part of the friable shonkinite where they are partly resorbed and zoned. Here, also, augite is most abundant, about 35 per cent of the rock, whereas in the typical, fine-grained shonkinite about 20 per cent is an average. The amount and kind of pyroxene change upward from the friable shonkinite to the syenite. Augite is practically absent in the syenite, and aegirite is common in all syenitic types. Aegirite-augite is common to the mafic syenite and top part

of the friable shonkinite. The cores of pyroxene are augite low in iron ($\beta = 1.688$); and the pleochroic, outer zones are richer in aegirite ($\beta = 1.725$). Some aegirite in pegmatitic cavities may be hydrothermal.

(3) In the rock of the chill facies, biotite is very fine-grained. Three varieties of biotite are present in other rocks: light-green, typical light yellow-brown to brown, and very pleochroic, red-brown. The red-brown variety is characteristic of the aplite syenites and is most commonly associated with aegirite in those rocks. The other two varieties are common in the shonkinites and mafic syenite. Where the green and yellow-brown varieties make up the same foil around olivine, the green variety is the older and grades into the brown. The distribution of the red-brown variety indicates that it is the youngest of the three. The mean refractive index measured on the cleavage flakes increases from 1.64 for the green biotite to 1.66 for the yellow-brown and higher for the red-brown biotite. There is likewise an increase in the intensity of the pleochroism. These optical variations are probably caused by the progressive increase in the amount of Fe_2O_3.

(4) Partly resorbed and altered crystals of vaguely but continuously zoned plagioclase are almost identical in the friable shonkinite, mafic syenite, and typical syenite. Its composition as indicated by its refractive indices is about An_{60} (sodic labradorite). Its size increases progressively upward from the base of the ribboned shonkinite. In the fine-grained shonkinite, the length of the tabular crystals is a small fraction of a millimeter, whereas in the friable shonkinite and syenite it averages between 1 and 2 millimeters. In the pegmatitic syenite the length is 3 millimeters. The grain size of the plagioclase is a good means of separating different units of shonkinite in the laccolith. For example, at least three layers make up the total thickness of the brim shonkinite west of station N. There is a sharp difference in the size of the plagioclase in the base of the brim shonkinite and in the immediately underlying ribboned shonkinite in many places. The absence of plagioclase in the analcime-rich aplite syenite dikes and dikelets is most significant and strongly suggests that those bodies were formed near the close of the magmatic stage of the central part of the laccolith.

(5) Fresh samples of sanidine in all the rocks are uniform in optical properties (2V about 45° (—), $\beta = 1.526$). Tabular crystals began to grow after the crystallization of the dark minerals and formed with such rapidity that they dominated the texture and compactness of the rocks. Tiny liquid (?) inclusions, arranged in rows and columns, are common in the coarse sanidine of the syenite. An altered variety of the sanidine, most common in the aplite syenites, is apparently homogeneous in plane polarized light but under crossed nicols has a "splotchy",

microperthitic (?) appearance. High magnification reveals a suggestion of albite twinning with lamellae having extinction angles differing by as much as 6 degrees. Interlocking, tabular crystals of sanidine have an average length in the syenite of about 4 millimeters, but crystals up to 3 centimeters long have been observed in pegmatitic syenite. Analcime is the chief alteration product of sanidine.

TABLE 3.—*Chemical analysis of analcime*

	Per cent of oxides	Molecular number
SiO_2	54.70	911
Al_2O_3	23.12	226
Fe_2O_3	.07	4
Na_2O	13.51	218
K_2O	.62	7
H_2O+	8.05	447
Total	100.07	

Formula: $(Na.K)_2O.Al_2O_3.4SiO_2.2H_2O$; or $NaAl(SiO_3)_2+H_2O$

(6) Of the hydrothermal minerals analcime is by far the most abundant, making up as much as one-fourth of some syenites. Specimens of this mineral are optically homogeneous. The mineral is sensibly birefracting, perceptibly twinned, biaxial (—), very small 2V, $\alpha = 1.488$, β and $\gamma = 1.489$ (all $\pm$ .001). It scratches glass easily. A large specimen of clear, glassy analcime was collected from a large ribbon on the face of the lower cliff west of station G, and its analysis by F. A. Gonyer is listed in Table 3.

(7) Crystals of apatite are not so abundant in the dikes of aplite syenite as they are in the adjoining mafic syenite. The mineral is most commonly associated spatially with the outer rims of aegirite-augite and in the foils of yellow-brown biotite.

(8) The calcite which formed along planes of fracture in the rubbled shonkinite has a pearly luster and such a flaky habit that the mineral appears to have a hardness of 1 to 2. It shows a dominant basal habit, is twinned, and, rarely, is biaxial with a small 2V. The softness of the calcite is probably caused by deposition in thin layers. Many grains which have the basal habit show a trigonal and hexagonal zonal development, probably the traces of rhombohedral growth. This flaky calcite is probably a high-temperature variety.

(9) Among the zeolites which have formed in cavities or areas interstitial to the tablets of feldspar, thomsonite is most abundant, and heulandite and natrolite are less abundant.

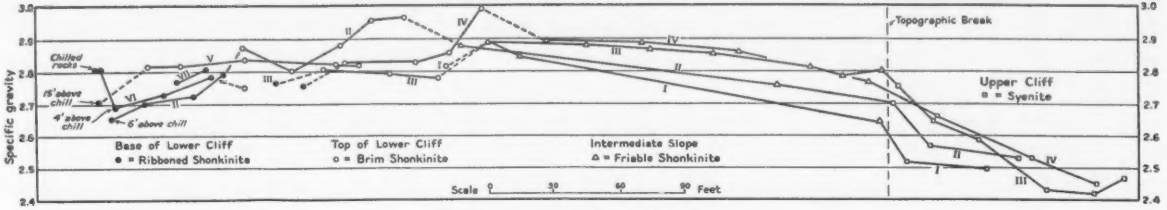

FIGURE 11.—*Specific gravity curves*

Composite chart showing plotted values of specific gravity for seven partial or complete sections on cliff faces of Boxelder Butte. Location of numbered sections are: (I) northeast of station P; (II) west of station G; (III) southeast of station L; (IV) west of station N; (V, VI, VII) partial sections east of station A.

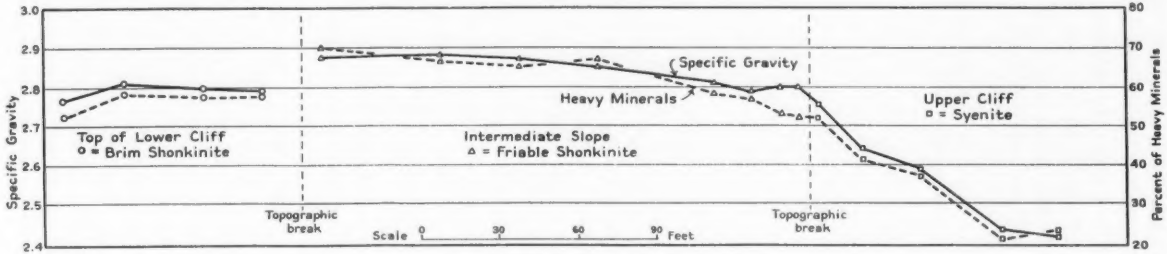

FIGURE 12.—*Heavy mineral curve*

Two curves of plotted values of specific gravity and heavy mineral percentage for rocks of section III in Figure 11.

(10) On the walls of many of the cavities in pegmatitic syenite, an unidentified mineral has formed as small bluish-white pellets attached chiefly on trapezohedra of analcime. A few of these pellets are attached to grains of magnetite, which in turn are attached to analcime. The pellets are soft and have a radiating, fibrous habit. In hand specimen, the pellets resemble the phosphate frankolite, but optical data and an X-ray powder picture exclude this identification. The mineral is probably a silicate. There is one good cleavage normal to Z, and the mineral is probably orthorhombic. The optical properties are: $\alpha = 1.611$, $\beta = 1.615$, and $\gamma = 1.626$, 2V medium (+), Y = elongation. Not enough material is available at the present time for a chemical analysis.

(11) In some cavities in pegmatitic syenite are small crystals of aegirite and of hornblende. The optical properties of the hornblende indicate an alkalic variety: α = brownish yellow, $\gamma = 1.725$, green, dispersion of the extinction position very strong.

QUANTITATIVE LITHOLOGIC RELATIONS

SPECIFIC GRAVITY VARIATIONS

Specimens were collected from many different places on the cliffs for the purpose of studying the exposed vertical section of the laccolith. The specific gravities were measured in the field with an improvised apparatus which gave a reproducible accuracy of ± 0.01 for the more compact specimens and ± 0.02 for the more friable ones. The measured densities of the porous rocks are undoubtedly a little less than their true values. In all, the specific gravities of more than 60 specimens were measured, and the values are plotted in the composite chart of Figure 11. Four complete and three partial sections of the igneous rocks are represented in the figure. Some specific gravities not shown on this chart but discussed in this paper are:

Aplite syenite dikes in mafic syenite	2.30
Aplite syenite dikelet in shonkinite	2.32
Pegmatitic plagioclase syenite	2.43

It should be noted here that several specimens of the chill facies of shonkinite have an average specific gravity of 2.81, and typical specimens of the brim shonkinite near the edge of the laccolith have an average value of 2.82.

HEAVY MINERAL VARIATIONS

The rocks of section III in Figure 11 were crushed to pass through a 100-mesh screen, and the heavy minerals were separated by means of bromoform and an electromagnet. The weight per cent of the heavy minerals has a probable accuracy of ± 2 per cent. In Figure 12 the per-

centage of heavy minerals in each rock is plotted with the value of specific gravity of the same rock. The two curves are comparable and should indicate that the values of specific gravity for other sections are sufficiently accurate for the purposes of this paper.

DISCREPANCIES IN DATA

The measured specific gravity of the friable shonkinite with 30 per cent of heavy minerals is 2.88, while the calculated density, assuming a fresh rock, is 3.03. This difference is probably due to porosity, the presence of analcime, and the alteration of olivine. The drop in the specific gravity of the shonkinite immediately above the chilled base of the laccolith (Fig. 11) is probably due to more complete alteration. The specific gravity of the brim shonkinite immediately below the friable shonkinite is 2.99 west of station N and 2.96 west of station G, higher than in the main part of the mass. This is due to a higher olivine content of the brim shonkinite.

The friable shonkinite and overlying mafic syenite near the center of the laccolith where the contact is gradational have higher values of specific gravity than rocks of the same horizons away from the center. The rocks near the center have more heavy minerals and also less analcime.

CALCULATED AVERAGES

As shown earlier in this paper, the syenite ribbons in two places constitute about 6 to 7 per cent of the ribboned shonkinite. The total amount of syenitic bodies in the friable shonkinite is estimated to be less than this, probably only a few per cent. It was concluded in the field and also from petrographic data that the brim shonkinite represents a younger intrusion than either ribboned or friable shonkinite and, further, that before the brim intrusions the ribboned shonkinite graded continuously up to the friable shonkinite. The transition in the field is marked by the great irregularity of the ribbons. Since the friable shonkinite has such an abundance of dark minerals and is not a separate intrusion, it is possible that this member of the laccolith was enriched in heavy minerals and impoverished in feldspathic minerals. Conversely, the syenite, as a member of the laccolith, was impoverished in heavy minerals. It is part of the thesis of this paper that both the friable shonkinite and the syenite are differentiated units and that their average bulk composition is comparable to undifferentiated shonkinite.

The weight per cent of heavy minerals in the chill facies of ribboned shonkinite is about 55, but this value is not so trustworthy as values for other rocks because of the fine grain size. The brim shonkinite with a heavy mineral percentage of 57 has a specific gravity of about 2.81 to

2.82. It is reasonable to assume that the percentage of heavy minerals in the chill shonkinite is therefore slightly less than 57.

The average value obtained for the heavy mineral percentages in the column of rocks shown in Figure 12 is 54 per cent. The method used was that commonly employed in mine evaluations, wherein values of the samples are weighted according to the distances between samples. This value of 54 per cent is low but within the range of accuracy of the value of heavy minerals for the undifferentiated shonkinite. The same method was used for the values of specific gravity in the same column of rocks, and the average was likewise comparable to the specific gravity of the chill facies of ribboned shonkinite.

RELATED SHONKINITIC BODIES

DIKES OF RIBBONED SHONKINITE

Numerous shonkinite dikes, varying in thickness from 3 to 12 feet, are intrusive into upper Cretaceous sediments in the plains area bordering the Bearpaw Mountains. The dikes are altered shonkinite much like that of the laccolith. All the dikes contain syenite ribbons, although many do not contain plagioclase.

About 4 miles southeast of the laccolith, 20 feet south of the Boxelder Creek road and one mile west of the Boxelder Creek dam, is the most interesting shonkinite dike in the mountains. The dike is vertical, well exposed, 7 feet thick, and strikes N. 50° W. Syenitic ribbons are conspicuous in the dike and they are thin and closely spaced near the walls but irregular and interconnected near the center where there are many small, irregular, coarse- and fine-grained segregation knots. The coarse-grained segregations grade texturally into the shonkinite.

DUCK CREEK LACCOLITH (?)

Five miles south of Boxelder Butte (Fig. 1) in the drainage area of Duck Creek is a butte with a cliff on the northern side and a summit which slopes gently to the south. About 100 feet of shonkinite characteristically ribboned with syenite caps the butte. The summit of the butte, however, is underlain by coarse shonkinite which contains many small and irregular syenitic bodies. The rock in general is more altered than that of Boxelder Butte but is petrographically similar. The butte near Duck Center is an erosional remnant of a larger body, probably similar in many respects to the Boxelder laccolith. On the southern side of the butte, sedimentary beds of the Judith River formation dip 20° S., and on the eastern side of the butte the shonkinite rests conformably on thick-bedded, buff sandstone of the same formation.

A careful examination of the summit of this butte for the purpose of finding a syenite plug proved fruitless. If such a plug ever existed, it would be more resistant than coarse shonkinite and topographically prominent. The summit is equivalent at least to the top part of the ribboned shonkinite in a comparison with Boxelder laccolith.

MOVEMENT OF MAGMA

FEEDER FOR LACCOLITH

No evidence of feeders has been observed for the Boxelder or Duck Creek laccoliths or for the several sills in the plains. There are no plug-like shonkinite bodies exposed in the area. The abundance of shonkinite dikes may suggest vertical migration of magma to a favorable horizon and lateral migration therefrom. Dikes have also been observed in the Fort Union sandstone at least 800 feet above the horizon stratigraphically equivalent to the Boxelder laccolith. It is possible that the upward advancing lobes of a few dikes reached a critical stratigraphic horizon by way of vertical fissures and spread along bedding planes.

CRITICAL HORIZON AND DEPTH OF COVER

The critical horizon for major lateral migration in the plains area on the western fringe of the Bearpaw Mountains appears to be the massive sandstone member near the top of the upper Cretaceous Judith River formation. Lithologic horizons in this formation cannot be correlated over distances of more than a few thousand feet, so that this massive sandstone is present only locally. Laccoliths or large sills have been observed only where relatively thick bedded sandstone is present. The critical horizon in the Bearpaw area is about 1000 to 1500 feet higher than the critical horizon for the Highwood Mountains (Eagle sandstone).

Dobbin and Erdmann (1934) have shown that the plains of Central Montana are underlain by a homoclinal structure plunging at a very low angle toward the east. According to them, the isolated mountain groups which rise above the level of the plains, as for example the Bearpaw and Highwood Mountains, have ruptured this homoclinal structure. Since their work, Pierce and Hunt (1937) have shown by structure contours on the Eagle sandstone that the region west of the Bearpaw Mountains is underlain by a broad and shallow syncline plunging eastward. The area between the two mountain groups (Pierce and Hunt, 1937, Pl. 43) is the south limb of this structure. Topographically, the Eagle sandstone near the Highwood Mountains and the Judith River sandstone near the Bearpaw Mountains are at the same general elevation. This structural situation, existing prior to the igneous intrusions of Tertiary time, can

satisfactorily explain the difference of the two critical horizons in the two mountain areas.

In the Bearpaw Mountains, above the Judith River formation, is exposed at least 1200 feet of sedimentary rocks of the Bearpaw shale, Lance sandstone, and Fort Union sandstone. If the formations were completely exposed, a thickness of about 2500 feet of sedimentary cover above the laccoliths is a reasonable estimate. In addition, a minimum thickness of 700 feet of andesitic and latitic flows, breccias, and conglomerates must be added to the thickness of the cover. At the time of shonkinitic intrusions, therefore, a minium depth of cover of 3000 feet over the Boxelder laccolith is certain, and this cover may have been twice as thick.

MECHANICS OF INTRUSION

Griggs (Hurlbut and Griggs, 1939) has discussed in some detail the mechanics of intrusion for the laccoliths of the Highwood Mountains. The field evidence observed in the Boxelder laccolith is in general support of his ideas.

There is plentiful evidence of some sort of forceful injection effecting deformation of the country rock at the time of intrusion. For example, in some places at the floor of the laccolith, particularly west of station G, minute folds and thrust faults with dimensions of a few inches have been induced in the thin-bedded sandstone immediately underlying shonkinite. In other places, large sandstone blocks have been disturbed from normal relations, and shonkinite in contact with younger intrusions has been fractured (rubbled). This evidence seems to suggest a lateral shouldering push during intrusion.

The writer does not support the operation of these forces by movement of liquid in the strict sense. One is forced to assume, however, that during intrusion the frontal, upper, and lower borders of a sill-like body would crystallize before the interior portion. There should be a stage in which the solid shell of the sill grades inward through crystalline mush to uncrystallized liquid, despite the rapidity of crystallization. If at this stage the intrusion is extended by a continuation of magma movement or by a second pulse of the intrusion, then the sill-like body as a unit might be caused to move outward, inducing a shouldering force. The rubbling in shonkinite described and illustrated in this report may have formed by this or a variation of this phenomenon. If the intrusive force is great, the magma will intrude the already solid frontal part of the sill. A series of pulsating surges would thus build up the steplike intrusions present at the base of the laccolith. In this manner, the roof of the laccolith could be constantly raised during the early

history of the laccolith and magma constantly intruded laterally at successively higher horizons.

The structures and form of the simple and multiple sills suggest rapid intrusion of a highly fluid magma. The fringing sills around the Boxelder laccolith are not so extensive as those around the Shonkin Sag laccolith, and this fact suggests that the two magmas, after lateral injection, did not have an equal fluidity. The greater amount of volatile material in the Boxelder magma would indicate that at the time of injection it was the more fluid; but a rapid loss of volatile material would then make the same magma more rapidly viscous and crystalline. The retention of volatiles in the central part of the laccolith would retard cooling.

PETROGENESIS

RIBBONS OF PLAGIOCLASE SYENITE

The syenite ribbons in shonkinite are a surficial expression of three-dimensional, tabular bodies representing a system of sheeting in shonkinite. The ribbons are generally present in the small shonkinitic bodies where there is a chilled border facies. The absence of ribbons in the brim shonkinite near the center of the laccolith and their presence in finer-grained brim shonkinite at the same horizon indicate that rapidity of crystallization is conducive to ribboning. The larger and interconnected ribbons about 30 to 40 feet above the chilled facies of ribboned shonkinite are much richer in analcime and other hydrothermal minerals than the smaller ribbons nearer the chilled facies. Perhaps a rapid loss of volatile material near the contacts aided in the rapidity of crystallization of the intrusion and in the formation of a well-defined pattern of parallel sheeting there. Because of less rapid crystallization away from the contacts, joints became less regular and fewer; and, because of the longer retention of volatile material within the body of an intrusion, larger ribbons with more analcime resulted.

The sheeting in the crystallizing shonkinite occurred after most of the dark minerals had formed and before much of the salic minerals had begun to form. From petrographic data a close temporal relation between sheeting and formation of the syenitic ribbons is imperative. It is likely that in the fine-grained shonkinite of small dikes and sills, where the ribbons have an orderly arrangement, the phenomenon of ribboning may have formed at nearly the same time throughout the body of shonkinite now exposed; but in the larger sills and dikes, as well as in the laccolith, may have occurred progressively inward from the contacts of the intrusions. As the sheeting planes developed, residual syenite soaked in along the parting surfaces. Some movement of syenitic

material may have occurred along the sheeting planes. The last-formed ribbons in the coarse-grained ribboned shonkinite may have formed after the plagioclase in the parent rock had crystallized, thus producing analcime-rich ribbons analogous to the syenite dikelets in the friable shonkinite.

The Bearpaw shonkinite is more altered than the described Highwood shonkinite and contains syenite ribbons which are lacking in the Highwood Mountains. It is most probable that a difference of volatile content of the two intrusive magmas has controlled the alteration and presence of ribboning in one and not the other. No evidence is available to indicate whether the difference in amount of volatiles (largely water) developed at depth or after irruption from depth upward through the sedimentary rocks.

SEGREGATION PLAGIOCLASE SYENITE

The shreds and patches of coarse-grained plagioclase syenite in coarse-grained plagioclase shonkinite or syenite appear to be true segregations of residual syenite. Little, if any, movement is postulated for the syenite of these bodies. They are generally coarser-grained than the rock in which they have formed and also contain more analcime. The more volatile-rich segregations developed a pegmatitic grain size and interstitial solution cavities lined with hydrothermal minerals. The characteristic femic minerals of the segregation syenite are aegirite and biotite.

DIKELETS OF PLAGIOCLASE-FREE SYENITE

The small, discontinuous dikelets of analcime-rich, plagioclase-free, fine-grained syenite in the friable shonkinite are probably a late, volatile-rich residual syenite. Some movement along joint planes in the friable shonkinite is postulated for the liquid which formed these bodies. Little, if any, movement is postulated for the small, pipe-like bodies, especially those in which skeletal plates of biotite have developed as shown in Plate 5. Since the ribboned shonkinite in some places is continuous into friable shonkinite, it is probable that the dikelets of plagioclase-free syenite are contemporaneous in part with the late-formed ribbons of analcime-rich syenite.

FRIABLE SHONKINITE

Much of the friable shonkinite contains appreciably more femic minerals than typical shonkinite. Among other data, the increasing amount of these minerals downward from the top can best be explained by a crystal settling process. The small volume of syenite in the friable shonkinite, either as groundmass or as individual bodies, suggests that some residual syenitic liquid was displaced by the settling crystals. The gradation of ribboned to friable shonkinite inward from the

border and upward from the base indicates that this fractionation took place within the central part of the laccolith insulated by a shell of ribboned shonkinite. Some lamellar flow and eddies may have been active during fractionation.

Although there is a relative increase in the amount of olivine from the top to the bottom of friable shonkinite, the total amount appears to be less than for an equivalent section of undifferentiated shonkinite. Also, there is more biotite in the friable shonkinite. The partial resorption of olivine in the laccolithic chamber may have supplied the crystallizing and fractionating shonkinite with ingredients later used in the formation of biotite. Petrographic data have proved that much of the biotite is later than the resorption of olivine.

The amount of heavy minerals in the lower part of the friable shonkinite is about 67 per cent, and this is approximately the amount found in the enriched parts of the laccoliths of the Highwood Mountains. It is believed that settling crystals will pack only to a concentration of about 70 per cent. In the thick Square Butte laccolith (?) the main shonkinite contains 72 per cent of dark minerals.

SIGNIFICANCE OF BRIM SHONKINITE

The brim shonkinite has been intruded into the laccolith at the zone of transition between the ribboned and friable shonkinite. Near the edge of the laccolith, the overlying friable shonkinite has a higher density than the exposed top of the brim shonkinite; but near the center of the laccolith, as for example near station G, the top of the brim shonkinite is composed of rock with a higher density than overlying friable shonkinite. The rock at the bottom of the brim shonkinite in both places is essentially the same. One explanation in keeping with the facts expressed above is that near the center of the laccolith a thin layer of mafic friable shonkinite contaminated the younger intrusion of brim shonkinite magma and thus caused a brim shonkinite with twice as much olivine as the undifferentiated, typical brim shonkinite elsewhere.

The brim shonkinite is most significant in that it represents a resurge or normal Boxelder shonkinite from depth after differentiation in the laccolith had begun.

PLAGIOCLASE SYENITE

The plagioclase syenite which caps Boxelder Butte is interpreted as the fractionated complement of the friable shonkinite. The chief evidence which leads to this conclusion is as follows:

(1) The friable shonkinite is not a separate intrusion.

(2) Differentiation has taken place within the friable shonkinite to produce small syenitic bodies.

(3) Large-scale differentiation by crystal settling has produced the friable shonkinite which is more mafic than undifferentiated shonkinite near the chill facies.

(4) In sections near the center of the laccolith the average density and heavy mineral percentage of both the friable shonkinite and syenite approximate the values for undifferentiated shonkinite.

(5) The friable shonkinite grades upward through mafic syenite to syenite near the center of the laccolith.

(6) The plagioclase is of the same character in the syenite and friable shonkinite.

(7) The olivine observed is of the same character in both the mafic syenite and friable shonkinite.

(8) Augite is abundant at the bottom of the friable shonkinite and grades upward to rare in the syenite. Aegirite is common in the syenite and grades downward to rare in the friable shonkinite. Aegirite-augite is common in both rocks but increases in proportion upward.

The field and laboratory evidence shows that it is possible for the plagioclase syenite as an individual member of the laccolith to have formed by fractionation of normal Boxelder shonkinite in place. Autoinjection may have occurred near the edge.

If, on the other hand, the alternative hypothesis of separate injection (Barksdale, 1937) is considered as an origin for the syenite, then two sources for a syenitic magma can be possible. One source is a derivative of fractionating shonkinite at depth; but in this event crystallization and differentiation at depth must have proceeded at a faster rate than that within the Boxelder laccolith, for it has been shown earlier that the brim shonkinite represents a recurrence of irrupting shonkinite after differentiation of the friable shonkinite had been effected. It is unlikely that fractionation at depth could have been rapid enough to allow a syenite magma to be intruded into the Boxelder laccolith while its center was still a partly crystallized liquid or a weak, hot, crystallized mush. It is strange, further, that, despite the many shonkinitic bodies in the plains area, syenitic plugs or dikes should be absent.

A second source of syenite magma can be derived by a special process of "bleeding" from the feeder dike or dikes of the laccolith the syenitic liquid which would have formed syenitic ribbons. This hypothesis is not, however, applicable to the Highwood laccoliths.

Both alternative hypotheses require a too special set of conditions of time and quantity of material and cannot account for other data. They require an introduction into the laccolith of a quantity of syenite

additional to that which would be derived by fractionation to form the friable shonkinite.

DIKES OF PLAGIOCLASE-FREE SYENITE

In the main body of residual syenitic liquid within the Boxelder laccolith, crystallization of aegirite-augite, aegirite, biotite, plagioclase, and

TABLE 4.—*Petrogenetic relations of the syenitic rocks to the main phases of crystallization of the Boxelder plagioclase shonkinite*

	Rock types	Olivine-augite phase →	Plagioclase-sanidine phase →	Sanidine-analcime phase
1a	Fine-grained shonkinite	X	X	X
b	Early-formed ribbons in fine-grained shonkinite		X	X
c	Late-formed ribbons in coarse-grained shonkinite			X
2a	Friable shonkinite	X	X	X
b	Segregation syenite		X	X
c	Dikelets and pipes in shonkinite			X
3a	Main syenite		X	X
b	Pegmatitic syenite		X	X
c	Dikes in syenite			X

sanidine left a further syenitic liquid which became progressively more volatile-rich. This liquid was injected along the sheeting planes developed in the mafic syenite and in a few places it was also injected downward into the friable shonkinite.

In the writer's opinion these dikes bear the same relation in petrogenesis (*see* Table 4) to the main syenite as the dikelets in the friable shonkinite bear to the groundmass or segregation syenite and as the last-formed analcime-rich ribbons in the top of the ribboned shonkinite bear to the early-formed ribbons of plagioclase syenite.

HYDROTHERMAL MINERALIZATION

The role of water was important in all stages of the magmatic and postmagmatic history. The quantity of water reduced the viscosity of the fractionated syenite, aided in the production of pegmatitic grain size, and caused the late magmatic autohydrothermal alterations. Calcite and zeolites are abundant products of hydrothermal mineralization, but quartz

is absent. The passage of truly magmatic to hydrothermal stage in the crystallization of the shonkinites and syenites may be regarded as reached when sanidine ceased to form and when analcime began to form. In light of this conclusion, it appears that some aegirite and hornblende in small solution cavities may be hydrothermal.

CONSTRUCTION OF THE LACCOLITH

The Boxelder laccolith is both a compound and multiple laccolith and was constructed in three principal stages: (1) First wave of shonkinitic intrusions, (2) Crystallization and differentiation, and (3) Second wave of shonkinitic intrusions. The beginning of the second stage was contemporaneous in part with the close of the first stage, and the beginning of the third stage may have been contemporaneous with the close of the second. Each of the three stages occurred rapidly.

The first wave of intrusion may have acted continuously or may have taken place as closely timed pulses. The initial, sill-like body which formed a nucleus for the laccolith was enlarged constantly by a raising of the roof and continual lateral injections at succeedingly higher horizons. Crystallization occurred immediately in the sills which fringed the main laccolithic chamber, and at the base and border of the laccolith. The earliest sills were fractured and partly displaced.

The second stage began with crystallization of shonkinite and production of the syenitic ribbons. The phenomenon of ribboning occurred continuously as the laccolithic reservoir was being built up. Upon temporary cessation of intrusion, cooling and crystallization encroached inward upon the still-liquid part of the laccolith from the border, base, and top of the laccolith. Hurlbut (1939) has described a reasonable mode of crystallization of the Shonkin Sag laccolith, and his views apply in general to the Boxelder laccolith. The magma in the liquid part of the laccolith held in suspension crystals of olivine and early-formed pyroxene. These crystals began to settle and were partly resorbed en route. The gradual subtraction of femic minerals left an increasingly syenitic residue which later crystallized as an individual member of the laccolith. Thus, as a result of the second stage, a compound shonkinite-syenite laccolith (*see* Fig. 13) was formed, and a representative curve showing variation of specific gravity for a vertical section is indicated by figure 14.

The third stage began with another wave of shonkinite (brim) and it spread out on a horizon near the bottom of the laccolith. The older shonkinite of the laccolith had crystallized and differentiated at the time of the second wave, but the rocks near the center were still hot enough to prevent the brim shonkinite from chilling there. Some contamination

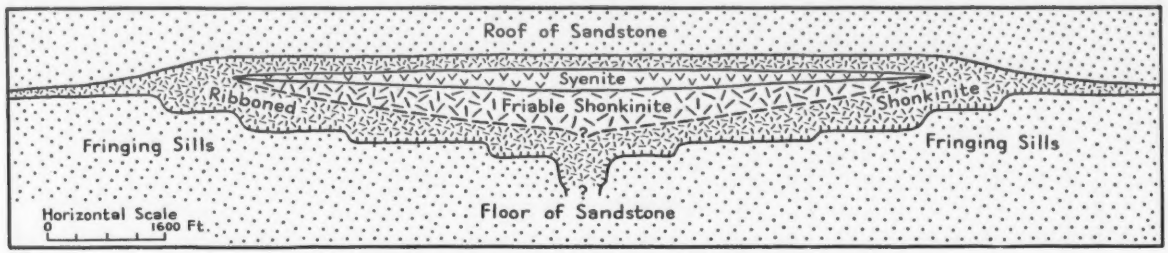

FIGURE 13.—*Idealized section of laccolith*

Prior to irruption of brim shonkinite (vertical scale exaggerated). Kernel of laccolith composed of differentiated shonkinite and syenite is enclosed in a shell of ribboned shonkinite.

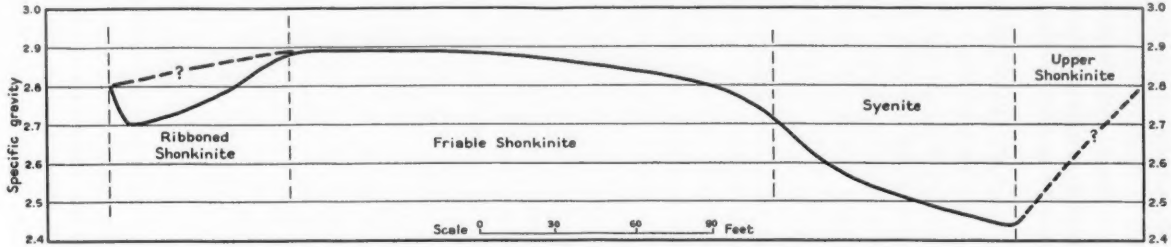

FIGURE 14.—*Specific gravity curve for idealized section*

Showing variation of values of specific gravity near center of Boxelder compound laccolith (prior to irruption of brim shonkinite). Broken line at right of figure represents interpreted upper shonkinite, and broken line at left represents theoretical values for unaltered ribboned shonkinite.

of this shonkinite may have been effected near the central part of the laccolith. Enough shonkinite was intruded in the second wave to be injected beyond the border of the already formed compound laccolith.

CONCLUSIONS

The Boxelder laccolith on the western fringe of the Bearpaw Mountains is a complex laccolith of the Shonkin Sag type which differs structurally and petrologically from any described laccolith in the Highwood Mountains. Field and laboratory data support the conclusion that the syenitic rocks differentiated from shonkinite after implacement of a shonkinitic magma into the laccolithic horizon. The progression of differentiation is from plagioclase shonkinite toward plagioclase syenite toward sanidine syenite.

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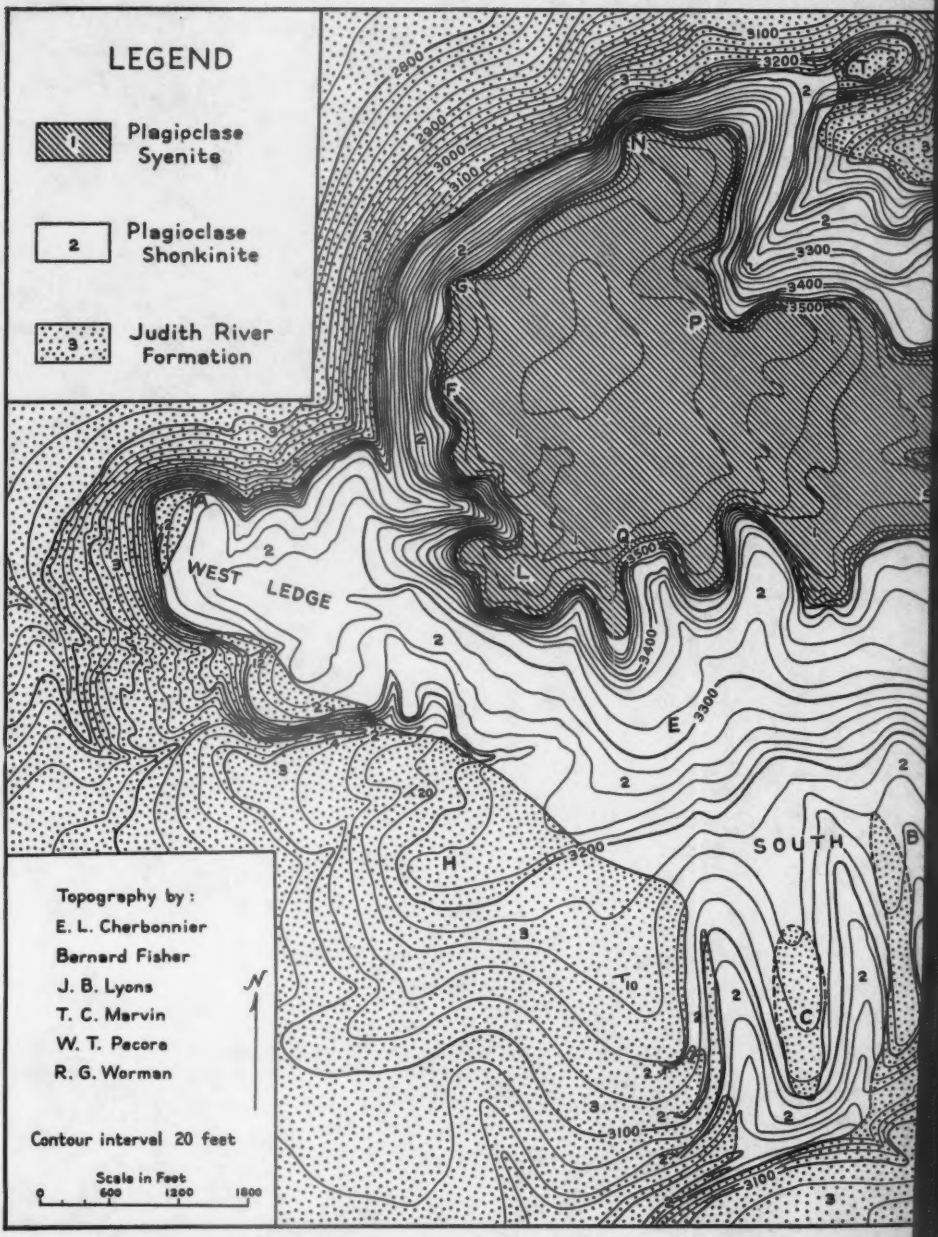
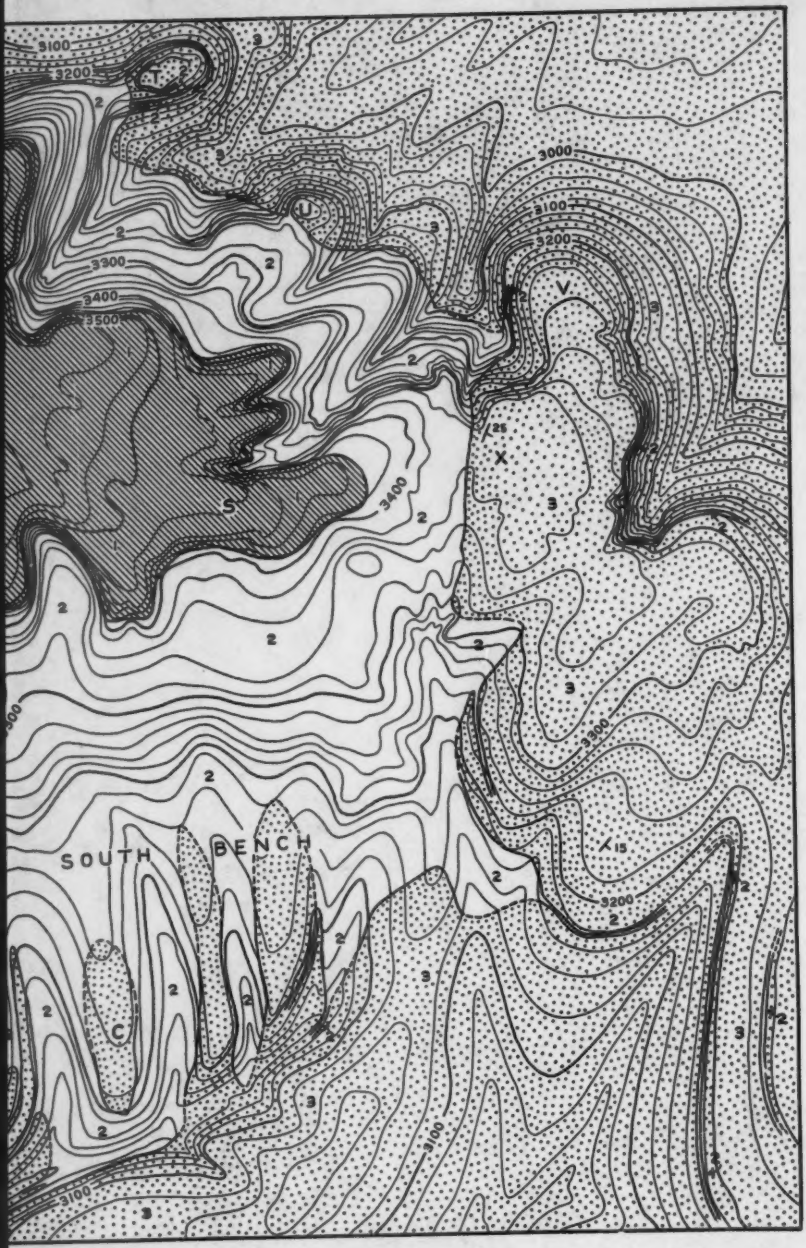


TOPOGRAPHIC AND GEOLOGIC MAP OF BOXE



LOGIC MAP OF BOXELDER BUTTE

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TRANSPORTATION OF ROCK BY KELP*

BY K. O. EMERY AND R. H. TSCHUDY

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ABSTRACT

In addition to being rafted by ice and by drifting logs, pebbles may be carried by kelp holdfasts to be dropped into deep-water sediments after the floating kelp has decayed or been eaten. Many shallow-water organisms similarly may be deposited in foreign environments. Numerous pebbles and cobbles still attached to kelp have been picked up on beaches, and in at least two instances pebbles with remnants of holdfasts have been found in relatively deep dredgings of the ocean floor. Although large kelps with floats are limited essentially to the Pacific and Antarctic regions, smaller algae from other coasts are also known to raft pebbles and shells long distances.

INTRODUCTION

In the past, geologists have occasionally noted the sporadic presence of pebbles and cobbles in fine-grained sedimentary rocks. Since water cur-

* Assistance is acknowledged from W. P. A. Project 65-1-07-2317.

rents which carry fine debris in suspension usually can not transport pebbles, some other agency is required to account for their occurrence in relatively well sorted sediments. In some instances the presence of these pebbles has been explained by their being rafted on icebergs or floes (Savage and Griffin, 1928; Reynolds, 1928; Bradley, 1940) and in others by their being carried in the roots of drifting logs (Spencer, 1887; Dana, 1895). More rarely suggested is transport by kelp holdfasts (Kinahan, 1879; Ball, 1888; Shaler, 1894; Ellis and Lee, 1919). This method of rafting is difficult to recognize in ancient sediments because the holdfasts are readily decomposed. However, examination of rocks gathered on beaches or recovered by dredging the sea floor throws more light on this means of transportation, since the rocks may be collected before the holdfast has been decomposed or eaten by organisms.

NATURE OF KELP

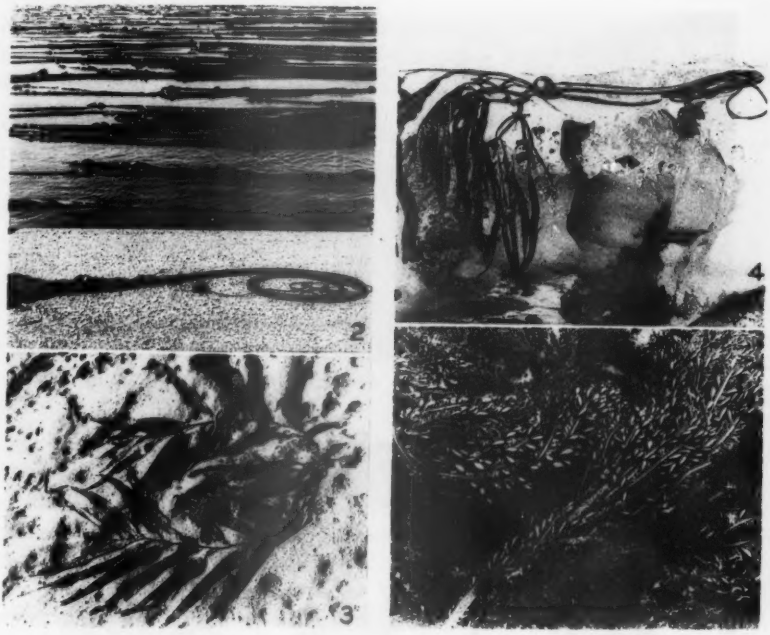
DESCRIPTION

General.—The term kelp has been used by some workers to designate the larger seaweeds. More recently, it has come to be used solely in reference to larger genera of the Laminariales, a family of the brown algae. With few exceptions, the larger seaweeds fall into this category.

Morphology.—The plant consists of two main parts—a holdfast, or rootlike structure, to anchor it to the bottom and an expanded photosynthetic portion called the blade. In some kelps the section between the holdfast and the blade, or lamina, is elongated and forms a long ropelike connection, the stipe.

The holdfasts of different genera vary considerably but usually they are made up of dichotomously branching structures which form a hemispherical mass anchoring the algae to the bottom. In growth, when the tips of these branches come in contact with a solid substratum, they spread out slightly and fasten themselves firmly to it. Usually the holdfasts do not exceed 12 inches in diameter, but in *Macrocystis* they have been known to reach a diameter of 3 feet.

Algal floats.—Some species of marine algae have distinct floats which hold the photosynthetic tissue close to the surface of the water. *Nereocystis* and *Pelagophycus* are the most spectacular representatives of this group (Pl. 1, figs. 1, 2, 4). The cavity of *Pelagophycus* floats from the kelp beds near La Jolla have a volume of about 120 cubic inches. Larger volumes are reported for floats of *Nereocystis* (Zeller and Neikirk, 1915). *Macrocystis* and *Egregia* have much smaller but very numerous floats (Pl. 1, figs. 3, 5). A single *Macrocystis* holdfast may have attached to it a dozen or more stipes, each of which possesses a float at intervals of about



LARGER FLOAT-BEARING GENERA OF BROWN ALGAE

(1) Kelp bed made up of *Nereocystis* plants. ($x\frac{1}{15}$).

(2) Individual *Nereocystis* plant. ($x\frac{1}{45}$).

(3) Young *Macrocystis* plant.
Showing arrangement of fronds and new branches arising from base. ($x\frac{1}{12}$).

(4) An individual *Pelagophycus* plant. ($x\frac{1}{60}$).

(5) *Egregia* exposed at extreme low tide. ($x\frac{1}{12}$).

ROCKS TRANSPORTED BY KELP

(1) Two small cobbles of sandstone and one of metavolcanic attached to a young *Macrocystis* holdfast. (x⅓).

(2) Kelp-rafted pebbles of cherty limestone.
Found in dredgings made at a depth of 720 feet. Each holdfast is *Egregia*. (x⅓).

(3) Shale; worm and pholad borings; *Macrocystis* holdfast. (x¼).

(4) Shale; worm and pholad borings; encrusting algae, Bryozoa, and so forth; *Pelagophycus* holdfast. (x¼).

(5) Granodiorite heavily encrusted by calcareous algae; *Macrocystis* holdfast. (x¼).

(6) Metavolcanic; calcareous algae, young *Macrocystis* holdfast. (x¼).

(7) Sandstone freshly broken from a larger mass; calcareous red algae on former exposed surface. (x¼).

(8) Two angular rock fragments weighing 4.5 pounds attached to a single *Pelagophycus* holdfast. (x¼).

6 inches along its entire length. The buoyancy of an entire *Macrocystis* plant is, therefore, great, even though each float has a volume of not more than half a cubic inch.

Size.—*Macrocystis* is the largest of the marine algae, commonly reaching a length of 200 feet and, exceptionally, as much as 600 feet. Adult plants of *Pelagophycus* and *Nereocystis* often are as much as 100 feet long, while *Egregia* and *Alaria*, the only other marine algae of comparable size, are usually not longer than about 60 feet.

The weight of a single *Macrocystis* plant has been known to be as much as 300 pounds, while other species commonly weigh not more than about 75 pounds (Frye, Rigg, Crandall, 1915).

Other much smaller species belonging to the family Laminariales play a definite though probably a relatively minor role in the transport of pebbles.

Distribution.—The kelps find their best development in the North Pacific Ocean and in northern European waters (Setchell, 1893). They are present along the west coast of North America from Alaska to south of Mexico and along the east coast of Asia, in the Okhotsk Sea, and off Japan. Skipping the equatorial region, they are again found along the west coast of South America, around the Antarctic continent, near Australia, and to a limited extent along the west coast of Africa.

The depth at which various algae are found depends upon several factors. The amount of light, turbidity of the water, and nature of the bottom are of importance. The brown algae are very seldom found growing in water as deep as 100 feet. The large kelps, such as *Macrocystis*, *Pelagophycus*, and *Nereocystis*, commonly occur in water from 20 to 60 feet deep. *Egregia* usually lives in somewhat shallower water. Many of the smaller genera of marine algae, such as *Eisenia*, *Lessoniopsis*, *Laminaria*, and *Pterygophora*, occur exclusively in the tidal zone or just below mean low water.

Geological history.—Although some kinds of algae are known from pre-Cambrian rocks, the earliest occurrences of supposed brown algae, insofar as the writers are aware, are those reported from the Devonian and Silurian. Species of *Nematophyton* reported by Lang (1926) show the filamentous internal structure of the stem believed by some to indicate a relationship to the Laminariales. More recently, Krausel and Weyland (1930) have discovered in rocks of these ages indefinite algae-like frond impressions connected with stems of the *Nematophyton* type.

An occurrence of two fossil brown algae from Miocene diatomite deposits has been reported by Gardner (1923). These algae appear remarkably similar to species living at the present time.

RESTRICTION TO ROCKY BOTTOM

Because of the pull exerted by strong currents, kelp must be securely anchored to the bottom. Good anchorage can be found only in places where hard ledge rock or large boulders are present. Reports on the extent of kelp beds along the west coast of the United States and Mexico show that they occur chiefly in rocky areas and practically never on sandy bottom, unless the sand constitutes only a shifting veneer over the rocks (Crandall, 1915; Brandt, 1923). Similarly, examination of United States Coast and Geodetic Survey charts shows notations of rocky bottom in areas of major kelp beds.

DESTRUCTION OF KELP BEDS BY LARGE WAVES

High waves developed by storms have been known to reduce kelp beds to about half their former size in a relatively short time. This depletion occurs both by the breaking up of the plants and by their dragging anchor. The holdfasts may be torn loose from the substratum, or the rock anchor itself may be lifted up or pulled apart. If one plant is broken loose and drifts against its neighbor, the combined pull of both may cause the second plant to break loose. Such a drive may tear a wide swath through a kelp bed, having been known to involve as many as 30 plants (Brandt, 1923).

KELP HOLDFASTS WASHED ON BEACHES

DEEPER WATER KELPS

From 1937 to 1940, holdfasts were occasionally picked up on the beach at the Scripps Institution of Oceanography at La Jolla, California. Although no systematic gathering was attempted, about 100 holdfasts were found largely by Dr. F. P. Shepard. These evidently were carried from a large kelp bed about a mile off shore across a submarine canyon more than 300 feet deep.

On June 5, 1940, moderately large waves carried a somewhat greater than usual number of kelps to beaches inshore from the same kelp bed. Distributed along about a third of a mile of the beach north of La Jolla, 46 *Macrocystis* and *Pelagophycus* holdfasts were found (Pl. 2), of which 17 had rock fragments attached. Along about half a mile of Pacific Beach, south of La Jolla, there were found 47 more holdfasts, 7 of which carried pebbles. Allowing for the kelp washed ashore on portions of the beaches not searched, it is probable that more than 100 holdfasts with pebbles attached were washed ashore after only one day of relatively large waves. Although it is impossible to determine the number of kelps washed out to sea, the finding of this number carried a

mile or more to the beach indicates the effectiveness of kelp in transporting pebbles. A great deal of kelp may be carried seaward from the mainland by storms with offshore winds and probably it is washed seaward continuously from kelp beds surrounding islands.

ALGAE FROM SHALLOWER WATER

In several instances, *Egregia* (Pl. 1, fig. 5), which lives in relatively shallow water near shore, has proven to be an important agent of rock transport. A cobble beach, south of La Jolla, was found to be abundantly covered by rocks which still had a portion of the algae attached. In an area of about 1200 square feet 110 cobbles were collected, of which 16 were attached to *Egregia* holdfasts, two to *Dictyota*, and one to *Neurocarpus*. Forty-six bore remnants of holdfasts which were too worn to permit identification of the type of alga. Forty-five other cobbles and pebbles were covered by encrusting organisms similar to those on the pebbles to which holdfasts were still attached. Since similarly transported rocks were nearly evenly distributed over the entire cobble beach, which is about 600 feet long and 75 feet wide, addition of rocks to the beach by kelp rafting over a long period of time must be substantial. That this condition is not unusual is supported by the collection of 207 holdfasts, chiefly of *Egregia*, on a cobble beach at Devil's Slide near La Jolla. Seventy-two had rock fragments attached, mostly of Chico, Cretaceous shale.

Although less significant than large kelps whose floats enable them to be carried a great distance, algae such as *Fucus, Dictyota, Neurocarpus, Eisenia, Lessoniopsis,* and *Pterygophora* cause the transport of pebbles a short distance toward shore from relatively shallow water (Johnson, 1919; Andree, 1920). Most of these algae possess no floats but are often torn loose from their anchorage by large waves at low tides and are carried to the beach. The force of the waves may be so great that the rocks to which the algae are attached may be transported to the beach along with the plant. After a period of large waves more than 100 plants of *Pterygophora* bearing rocks of various sizes were picked up along a few hundred feet of sandy beach at Point Conception, California.

That rafting by these shallow-water algae is not restricted to short distances is shown by Collins (1914), who reports finding pebbles and shells in the holdfasts of *Laminaria, Fucus, Desmarestia, Ascophyllum,* and *Rhodymenia* which had floated 150 miles from their point of origin.

CHARACTER OF ATTACHED PEBBLES

About two-thirds of the rock fragments picked up on beaches inshore from the La Jolla kelp beds consist of shale or sandstone probably of

Chico, Cretaceous and Rose Canyon, Eocene age. Most of the other specimens are rounded cobbles of Black Mountain meta-volcanics which occur in Eocene conglomerates. Some chlorite schists of unknown derivation were also found. All the rocks are partially encrusted by calcium carbonate depositing organisms such as Bryozoa, Mollusca, serpulid worms, and calcareous red algae (Pl. 2). Worms and pholads have made deep borings in the softer rocks.

The largest rock obtained is about 15 inches long, and the heaviest weighs about 13 pounds. Most of the holdfasts are quite large, though the pebbles attached are small. In one instance, more than 25,000 pebbles and granules larger than one millimeter in diameter were held by one holdfast. This suggests that a holdfast may be wrenched loose from a solid substratum but may retain its grasp on small loose pebbles which had lain on the surface of a larger rock mass.

TRANSPORT OF ORGANISMS

Many holdfasts picked up on beaches or found drifting in the water have shells attached to them. Fragments of pelecypod and gastropod shells are occasionally found, together with the more common serpulid worm tubes. Living organisms include starfish, brittle stars, echinoids, sponges, and many types of worms (Andrews, 1938). Encrusting organisms such as Bryozoa, Foraminifera, worm tubes, calcareous red algae, and others are attached to rocks of the holdfast (Pl. 2). Similar instances of transport of organisms by algae have been noted by Vallentin (1895). All these organisms are carried from their normal environment and are dropped by the floating kelp in areas where they would not normally live, perhaps to be incorporated in beach sands or in sediments of the deep sea.

HOLDFASTS ENCOUNTERED IN DEEP-WATER DREDGINGS

During 1938, many rock dredgings of the ocean floor off southern California were made from the research vessel E. W. Scripps. Samples at two localities include pebbles having an attached kelp holdfast.

At the head of Pedro Trough (a submarine canyon about 5 miles south of the Palos Verdes Hills) one dredging recovered an angular pebble 2 inches long, to which part of a holdfast was still fastened (Pl. 2, fig. 2). The rock is cherty limestone, possibly of Miocene age. Another dredging, taken half a mile away, also contained a holdfast and an attached pebble. This rock is smaller than that of the first dredging but is of similar lithology. Both holdfasts were found to be of *Egregia*. These dredgings were made at a depth of 720 feet, far below the depth of kelp growth. In each instance, the kelp probably floated seaward from the extensive beds surrounding the Palos Verdes Hills.

One questionable kelp-rafted pebble was found in another sample from the ocean floor. A core, taken in clean sand at a depth of 180 feet in a submarine canyon off La Jolla, contained a rounded pebble of Black Mountain metavolcanic rock similar to those frequently found attached to holdfasts picked up on the beach near-by. This pebble has what appears to be a holdfast, but so little remains that it cannot be positively identified as such. Farther from shore in the same canyon a dredging at 1320 feet recovered a holdfast, but no rock was attached.

SIGNIFICANCE OF KELP TRANSPORT

Since kelp holdfasts containing pebbles have been found frequently on marine beaches and also are known from dredgings of the sea floor, this method of transport should receive more attention than has been accorded it. Such an agency may explain the occasional finding of erratic pebbles in otherwise uniform fine-grained sediments. The presence of borings and of encrusting organisms on such erratic pebbles may serve to strengthen the possibility of kelp rafting. However, it is likely that most of the rocks carried by kelp are deposited on beaches and are reworked into the beach conglomerate. Abrasion through the action of beach processes quickly removes all traces of surface encrustations, making the identification of kelp-rafted pebbles in ancient conglomerates difficult or impossible. However, pebbles of unique lithology, angularity, or size with respect to others in a conglomerate may have been carried by kelp to the beach.

Although only two certain cases of kelp-rafted pebbles have been noted in dredgings off California, it must be remembered that dredgings and cores have recovered rocks from only an infinitely small proportion of the total area of the ocean floor.

During cruises of the research schooner, E. W. Scripps, floating kelp has been noted as far as 250 miles from the nearest land (Roger Revelle, personal communication). Pebbles could easily have been carried by this floating kelp and have been dropped when enough of the floats had decayed or been eaten by organisms.

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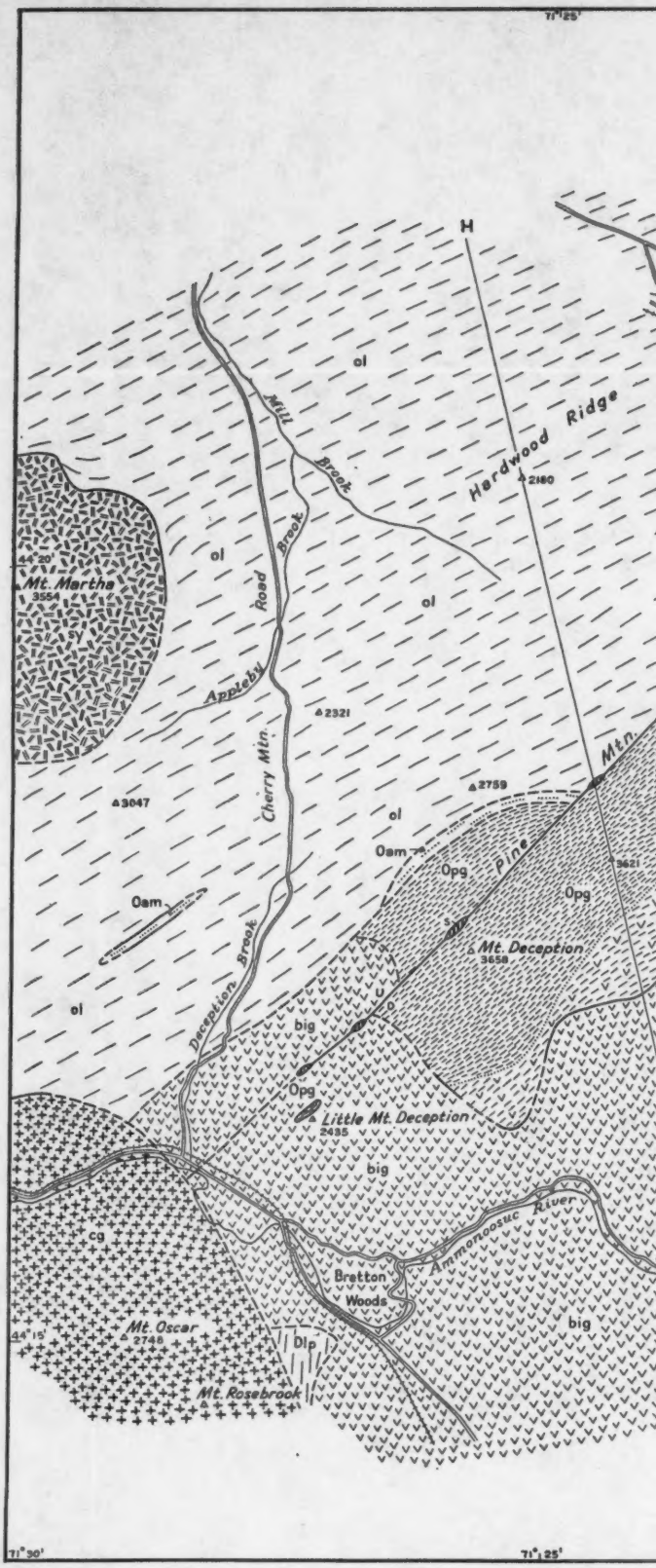
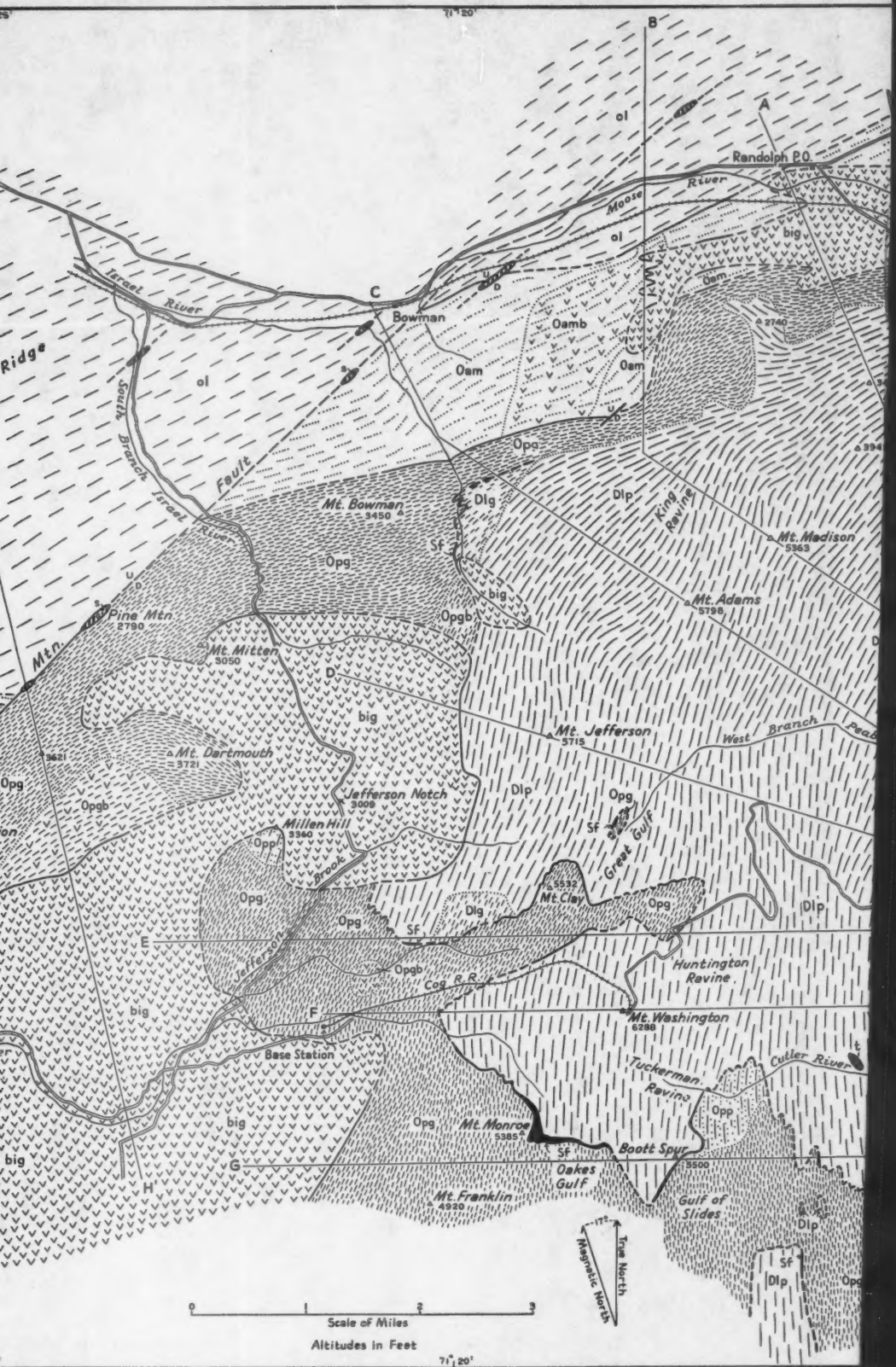
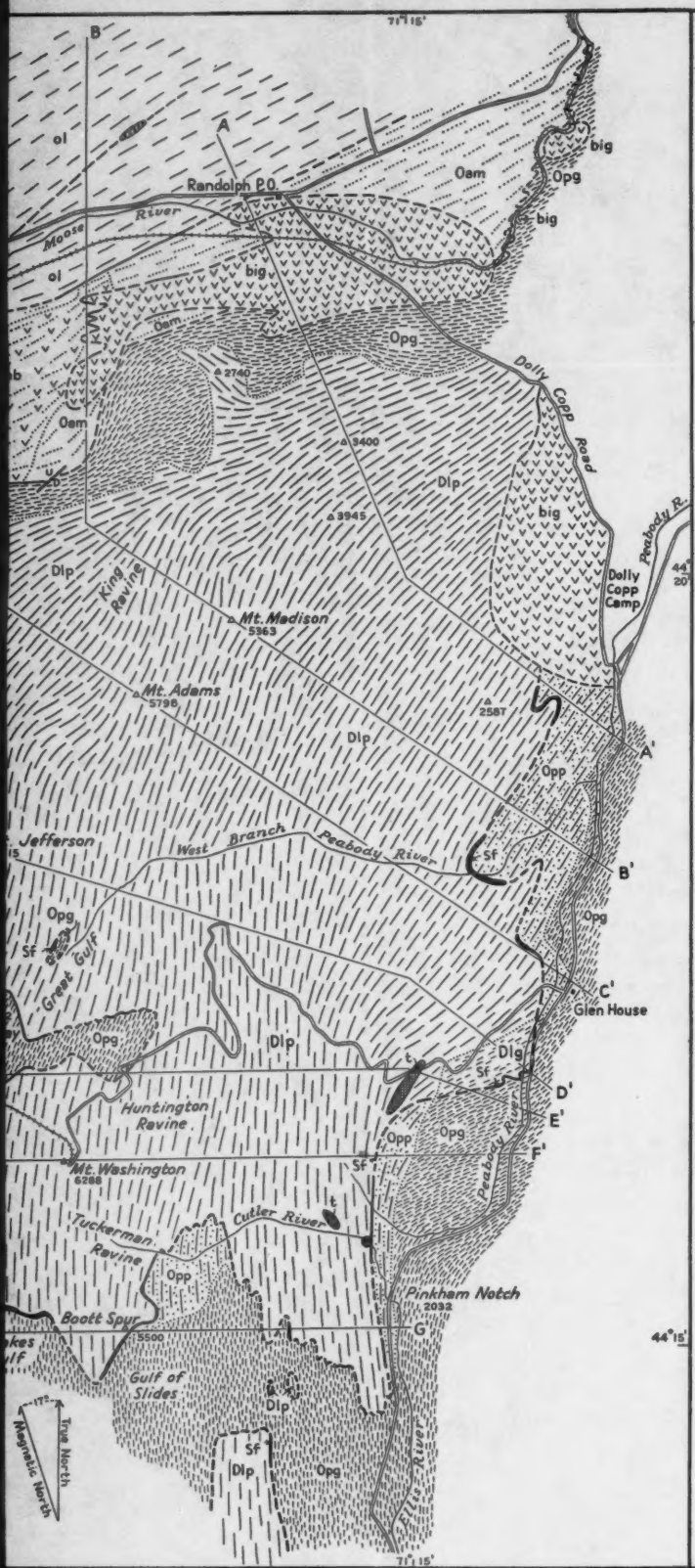


Linear patterns ind



GEOLOGIC MAP OF MT. WASHINGTON AREA

patterns indicate trend of axes of folds in metamorphic rocks and foliation in plutonic rocks. Nonlinear patterns



NGTON AREA

n plutonic rocks. Nonlinear patterns indicate massive rocks.

LEGEND

Mississippian ?

White Mountain Magma Series

cg — Conway granite

sy — Syenite

t — Tuff, breccia, and diabase in volcanic vents

Late Devonian ?

New Hampshire Magma Series

big — Bickford (=Concord ?) granite

Oliverian Magma Series

ol — (Various units not distinguished on this map)

Lower Devonian

Dlp, Dlg — Littleton formation
Dlp – Paraschists
Dlg – Gneiss

Middle Silurian

Sf — Fitch formation

Ordovician ?

Opp, Opg, Opgb — Partridge formation
Opp – Paraschists
Opg – Gneiss
Opgb – Gneiss with dikes of Bickford granite

Oam, Oamb — Ammonoosuc volcanics
Oamb = Ammonoosuc volcanics with many dikes and sills of Bickford granite

CONTACTS

——— Accurate
– – – – – Approximate
····· Transitional

Silicified zone

U / D — Normal fault
D = Downthrown side

Altitudes in feet

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STRUCTURE AND METAMORPHISM IN THE MOUNT WASHINGTON AREA, NEW HAMPSHIRE

BY MARLAND P. BILLINGS

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ABSTRACT

The Mt. Washington area of New Hampshire contains the highest peaks in the northern Appalachian Mountains. The metamorphic rocks belong to four stratigraphic units: the Ordovician (?) Ammonoosuc volcanics, chiefly fine-grained biotite gneiss; the Ordovician (?) Partridge formation, largely gneiss derived from shale; the Silurian Fitch formation, consisting of lime-silicate granulite and schist; the Devonian Littleton formation, mostly quartzite and schist.

The oldest intrusive rocks are quartz monzonites and granites belonging to the Devonian (?) Oliverian and New Hampshire series. Granite, syenite, tuff, and breccia belong to the Mississippian (?) White Mountains magma series.

The chemical composition of some of the minerals differs with the formation. Pyroxene and amphibole in the Fitch formation are magnesia-rich and iron-poor, whereas in the Ammonoosuc volcanics the reverse is true. Oligoclase characterizes the Partridge formation, Littleton formation, and Bickford granite; andesine is typical in the Ammonoosuc volcanics; and bytownite is common in the Fitch formation. Biotite and muscovite are relatively uniform, except for magnesia-rich biotite in the Fitch formation.

The Mt. Washington area is on the southeast flank of a large dome, the center of which is occupied by the intrusive Oliverian magma series. The folds in the Presidential Range trend north and northeast. The major folds are *en echelon*, and upon them are superimposed many minor folds. Schistosity, due to platy minerals, parallels the bedding. Fracture cleavage is essentially parallel to the axial planes of the minor folds. The Pine Mountain fault is the largest of several normal faults.

Metamorphism is high-grade (katazonal). Original shales are now andalusite schist, coarse rough pseudo-andalusite schist, fine grained pseudo-andalusite schist, and staurolite schist. Contrasting metamorphic history explains the various types. Metamorphism was syntectonic, and the coarser schists show three major stages in the paragenesis. Many rocks, particularly paraschists, suffered no significant chemical

change. Potash was introduced into some of the coarser schists. Many of the gneisses are derived from shale by metamorphic differentiation, and less than 1 per cent of soda, lime, and potash has been introduced. Four or five per cent of soda, lime, and potash has been added to shale to form the lighter colored gneisses.

INTRODUCTION

The Mt. Washington area, as defined for this paper, covers about 100 square miles somewhat north of the center of New Hampshire (Fig. 1). The mapped area is mostly in the southern half of the Mt. Washington quadrangle, but includes a few square miles in the southwestern part of the Gorham quadrangle and the northern part of the Crawford Notch quadrangle.

Brief mention of the geology was made by C. T. Jackson (1844) and H. D. and W. B. Rogers (1846; 1848). The most thorough investigation of the bedrock geology was by C. H. Hitchcock (1877, p. 113-127). R. W. Chapman (1937) has described the syenite stock on Cherry Mountain (Mt. Martha). No attempt was made in the present investigation to consider all phases of the geology. The study was concerned with the lithology, stratigraphy, structure, and metamorphism. The physiography and glaciation have been so thoroughly studied by J. W. Goldthwait (1916), A. C. Lane (1921), I. B. Crosby (1924), E. Antevs (1932), and R. P. Goldthwait (1940), that the present writer did no work on these subjects.

The field work occupied one month in the summer of 1936 and three months each during the summers of 1938 and 1939. On topographic maps, enlarged to about 3 inches to the mile, all observations were plotted, and a complete outcrop map was prepared. Location was made largely by barometer, inasmuch as all streams, ridges, and trails are shown on the map. A few pace-and-compass traverses were made but it was not necessary to use this method as extensively as in the Littleton-Moosilauke area (Billings, 1937, p. 466-469). The laboratory investigation was carried on during the winters from 1937 to 1940. Two hundred and thirty thin sections were used and a great deal of optical mineralogy was essential. Nine chemical analyses were made by the Rock Analysis Laboratory at the University of Minnesota.

After the field work had been completed, it became obvious that the results could be most satisfactorily presented in a series of papers. The dike rocks, although variable and interesting, have no particular relationship to the stratigraphy, metamorphism, and structure and will be considered in a separate paper by Katharine Fowler-Billings. The Oliverian magma series consists of several mappable units. However, these rocks are also extensively developed in the northern half of the Mt. Washington quadrangle, now being studied by R. W. and C. A. Chapman, and will be considered in a joint paper. The stock of syenite on

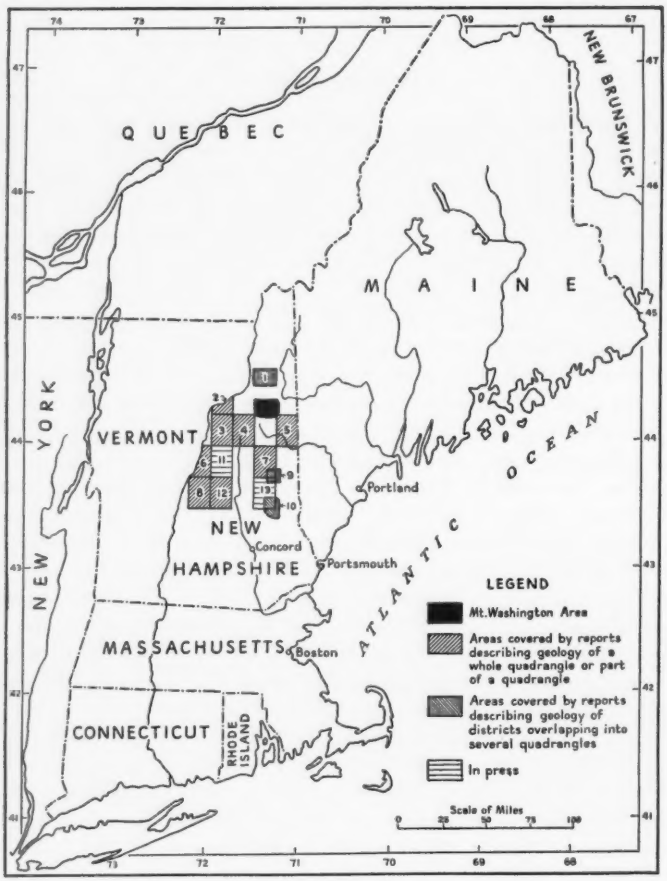


FIGURE 1.—*Index map*

Shows location of the Mt. Washington area in solid black. Other areas studied in the present investigation of New Hampshire are: 1 = Percy area (R. W. Chapman, 1935); 2 and 3 = Littleton and Moosilauke quadrangles (Billings, 1937); 4 =Franconia quadrangle (Williams and Billings, 1938); 5 = North Conway quadrangle (Billings, 1928); 6 and 8 = Mt. Cube and Mascoma quadrangles (Hadley and C. A. Chapman, 1939; C. A. Chapman, 1939); 7 = Mt. Chocorua quadrangle (Smith, Kingsley, and Quinn, 1939); 9 = Ossipee Mountains (Kingsley, 1931); 10 = Belknap Mountains (Modell, 1936); 11 = Rumney quadrangle (Page, in preparation); 12 = Cardigan quadrangle (Fowler-Lunn and Kingsley, 1937); 13 = Winnepesaukee quadrangle (Quinn, in preparation). Red Hill (Quinn, 1937) lies west of the Ossipee Mountains.

FIGURE 1. AERIAL VIEW FROM THE SOUTH IN WINTER
The highest peak is Mt. Washington. (Photo by H. Bradford Washburn, Jr.)

FIGURE 2. AERIAL VIEW FROM THE WEST
Peaks from left to right: A—Mt. Adams; J—Mt. Jefferson; C—Mt. Clay; W—Mt. Washington; B—Boott Spur; M—Mt. Monroe; P—Pleasant Dome, which is not within the area covered in Plate 1. Open fields in foreground are Bretton Woods. Lowland is underlain by Bickford granite. (Photo by H. Bradford Washburn, Jr. Courtesy of The Flume Reservation, Franconia Notch, New Hampshire)

FIGURE 1. AERIAL VIEW FROM THE EAST
B—Boott Spur; W—Mt. Washington; C—Mt. Clay; S—Gulf of the Slides; T—Tuckerman Ravine; H—Huntington Ravine; G—Great Gulf. (Photo by H. Bradford Washburn, Jr. Courtesy of the Flume Reservation, Franconia Notch, New Hampshire)

FIGURE 2. AERIAL VIEW OF TUCKERMAN RAVINE FROM THE EAST
Shows axis depression just above the headwall. To the left (south) of the Ravine the folds plunge north; above the headwall they are horizontal; and to the right they plunge south. (Photo by H. Bradford Washburn, Jr.)

MT. WASHINGTON

Cherry Mountain (Mt. Martha) has been described by R. W. Chapman (1937). The Conway granite in the southwestern part of the quadrangle is so similar to its continuation in the Franconia quadrangle (Williams and Billings, 1938) that a detailed description is unnecessary. Two preliminary papers on the area have already been published (Billings, 1938b; 1939).

ACKNOWLEDGMENTS

The writer is greatly indebted to numerous organizations and individuals for aid. The cost of the study was partially financed by grants from the Associates in Science of Harvard University and from the Penrose Bequest of The Geological Society of America. Robert Williams assisted in the field in 1936, F. B. Loomis, Jr. in 1938, and W. P. Fuller, Jr. and R. F. Story in 1939. Mr. Loomis mapped the area bounded on the south by Ammonoosuc River, on the west by the Cherry Mountain road, on the east by the Jefferson Notch road, and on the northwest by a northeast-southwest line 1 mile northwest of Pine Mountain. Katharine Fowler-Billings was my co-worker in the field throughout the summer of 1938. Mr. Fuller gave valuable laboratory assistance during the winter of 1939-40.

Mr. Elliott Libby, of Gorham, New Hampshire, kindly allowed us to use the automobile road on Mt. Washington without charge. Due to the great hurricane of September, 1938, large portions of the White Mountain National Forest were closed to all entry during the summer of 1939 because of great fire-hazard. Mr. C. L. Graham, Forest Supervisor, kindly granted permits to enter closed areas, and the district rangers, N. D. Shirley and H. C. Waldo were most cordial. Oblique aerial photographs by H. Bradford Washburn, Jr. proved useful in planning accessible routes and in locating outcrops. The original manuscript has been extensively revised in accordance with suggestions made by C. A. Chapman and R. W. Chapman. Photographs accompanying this paper, unless otherwise acknowledged, are by Katharine Fowler-Billings, R. F. Story, and W. P. Fuller, Jr. Edward Schmitz has done the drafting.

TOPOGRAPHY AND DRAINAGE

The Mt. Washington area (Pl. 1), containing the highest peaks in the northern Appalachian Mountains, has a maximum difference in elevation of somewhat greater than 5000 feet. The Presidential Range (Pls. 2, 3), which occupies the eastern half of Plate 1, trends north-south but is convex toward the west. The principal summits, from south to north, are: Boott Spur, Mt. Monroe, Mt. Washington (altitude 6288 feet), Mt. Clay, Mt. Jefferson, Mt. Adams, and Mt. Madison. To the east of the

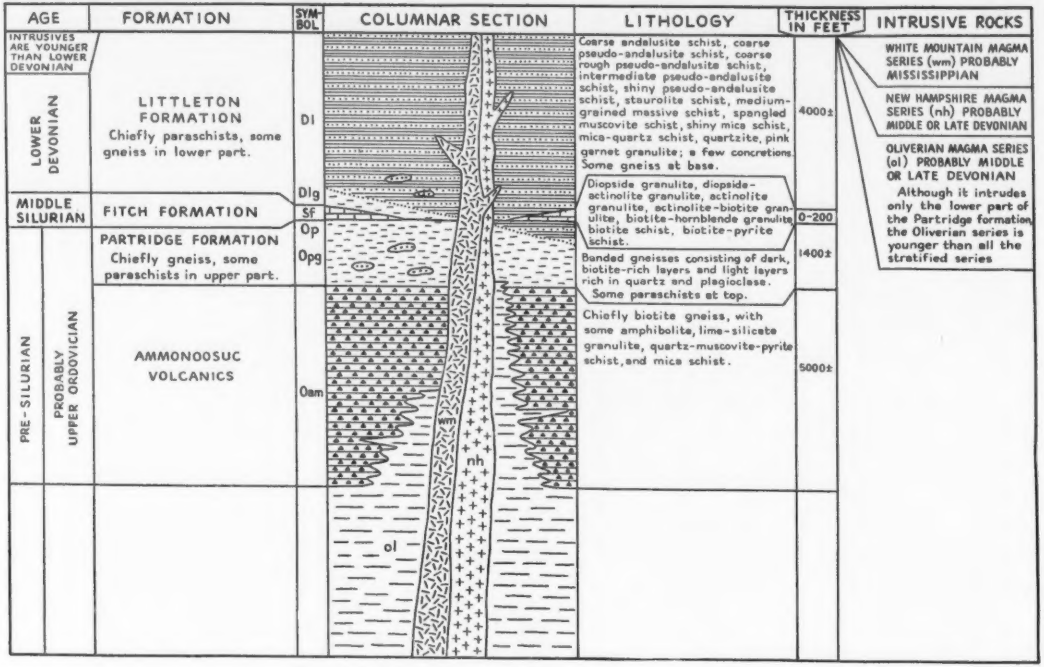

AGE
FORMATION
SYMBOL
COLUMNAR SECTION
LITHOLOGY
THICKNESS IN FEET
INTRUSIVE ROCKS
INTRUSIVES ARE YOUNGER THAN LOWER DEVONIAN
LOWER DEVONIAN
LITTLETON FORMATION
Chiefly paraschists, some gneiss in lower part.
Dl
Coarse andalusite schist, coarse pseudo-andalusite schist, coarse rough pseudo-andalusite schist, intermediate pseudo-andalusite schist, shiny pseudo-andalusite schist, staurolite schist, medium-grained massive schist, spangled muscovite schist, shiny mica schist, mica-quartz schist, quartzite, pink garnet granulite; a few concretions. Some gneiss at base.
4000±
WHITE MOUNTAIN MAGMA SERIES (wm) PROBABLY MISSISSIPPIAN
NEW HAMPSHIRE MAGMA SERIES (nh) PROBABLY MIDDLE OR LATE DEVONIAN
OLIVERIAN MAGMA SERIES (ol) PROBABLY MIDDLE OR LATE DEVONIAN
Although it intrudes only the lower part of the Partridge formation, the Oliverian series is younger than all the stratified series
MIDDLE SILURIAN
FITCH FORMATION
Dlg
SF
Diopside granulite, diopside-actinolite granulite, actinolite granulite, actinolite-biotite granulite, biotite-hornblende granulite, biotite schist, biotite-pyrite schist.
0-200
PARTRIDGE FORMATION
Chiefly gneiss, some paraschists in upper part.
Op
Opg
Banded gneisses consisting of dark, biotite-rich layers and light layers rich in quartz and plagioclase. Some paraschists at top.
1400±
PRE-SILURIAN
PROBABLY UPPER ORDOVICIAN
AMMONOOSUC VOLCANICS
Oam
Chiefly biotite gneiss, with some amphibolite, lime-silicate granulite, quartz-muscovite-pyrite schist, and mica schist.
5000±
wm
nh
ol

FIGURE 2.—*Columnar section for Mt. Washington area*

Shows sequence, character, and thickness of the metamorphic rocks and the sequence of igneous intrusion. The Oliverian magma series (ol) at the base of the column is not part of the stratified series, but a huge intrusive sheet, the top of which varies considerably in stratigraphic position.

Presidential Range are the Peabody and Ellis rivers, which flow northeast and south respectively. Pinkham Notch (altitude 2032 feet) is at the divide.

The Dartmouth Range lies in the central part of the area, trends northeast, and contains the following principal peaks: Millen Hill, Mt. Dartmouth (altitude 3721 feet), Mt. Deception, and Little Mt. Deception. Cherry Mtn., the principal summit of which is known as Mt. Martha, is in the west-central part of the area. Moose and Israel Rivers drain the northern part, and the Ammonoosuc River drains the southwestern part.

The area is heavily forested, except for the higher parts of the Presidential Range (Pl. 4, fig. 2). Timberline (Pl. 4, fig. 1) is at about 4500 to 4800 feet, although locally it is lower. Outcrops are not abundant above timberline, due to extensive late-glacial or postglacial frost action. Bedrock is well exposed in cirques, along many of the streams, on landslides, and in numerous outcrops scattered in the forest. As a whole, outcrops are fairly good, but are scarce on the northern slopes of the Dartmouth Range.

GENERAL LITHOLOGIC FEATURES

Metamorphic and plutonic rocks compose the Mt. Washington area. Their sequence, age, lithology, thickness, and a graphical representation of the geological history are given in Figure 2.

The oldest rocks, probably upper Ordovician, are the Ammonoosuc volcanics. Above is the Partridge formation, also probably upper Ordovician. It is largely gneiss, but locally is schist and quartzite. The middle Silurian Fitch formation is a thin, discontinuous unit above the Partridge formation. It has been derived from impure dolomites and is distinctive because it contains diopside, actinolite, calcic plagioclase, and magnesia-rich biotite. The highest stratigraphic unit is the lower Devonian Littleton formation, which is largely quartzite and schist. The lower part of the formation has been locally converted to gneiss. The total thickness of the metamorphosed sedimentary and volcanic rocks is about 10,500 feet.

These stratified rocks are intruded by plutonic rocks belonging to three of the four magma series of New Hampshire. The Oliverian magma series is younger than the stratified rocks but is slightly older than the orogeny, probably middle or late Devonian. The New Hampshire magma series is represented by the Bickford granite, probably injected just after the folding and thus late Devonian. The White Mountain magma series is represented by the syenite on Cherry Mountain (Mt. Martha), the Conway granite around Mt. Oscar, and small volcanic vents. This series is probably Mississippian.

Over 100 basic dikes, belonging partly to the White Mountain magma series and partly younger, have been observed, but will be described in a separate paper. Glacial drift hides much of the bedrock.

AMMONOOSUC VOLCANICS (ORDOVICIAN?)

GENERAL STATEMENT

Along the lower slopes of the northern flank of the Presidential Range the Ammonoosuc volcanics occupy a belt which trends east-northeast and is almost 1½ miles wide (Pl. 1). The rocks are well exposed along streams from south of Bowman to southwest of Randolph. Southwest of Mt. Bowman, however, exposures of the Ammonoosuc volcanics are rare. South of Randolph the belt of Ammonoosuc volcanics is interrupted by a stock of Bickford granite and by glacial drift. Isolated exposures are found, however, along the railroad 2 miles east of Randolph. The small belt of Ammonoosuc volcanics shown 1 mile north of Mt. Deception is based on float; no outcrops have been observed. The small belt 2½ miles west of Mt. Deception is well exposed, and the intrusive relations of the surrounding granites of the Oliverian magma series are well established.

The best exposures of the Ammonoosuc volcanics are along Israel River, Moose River, and several streams not shown on Plate 1, especially The Mystic, Cold Brook, and Snyder Brook.

LITHOLOGY

General character.—The Ammonoosuc volcanics are classified as follows: (1) amphibolite, including quartz amphibolite; (2) biotite gneiss, derived from tuffs with the composition of rhyolite, quartz latite, and dacite; (3) lime-silicate granulite and schist; (4) quartz-muscovite-pyrite schist; and (5) miscellaneous rocks of sedimentary origin, including mica schist, biotite-feldspar schist, and quartz-chlorite-plagioclase gneiss.

Amphibolite.—Amphibolites are not common in the Ammonoosuc volcanics of the Mt. Washington area. Megascopically they are dark, speckled, medium-grained rocks composed of black hornblende and white plagioclase. A weak schistosity is generally present, and in some cases the aligned hornblende needles give a distinct lineation.

Under the microscope the amphibolites are granoblastic; hornblende, 0.1 to 1.0 millimeters in diameter, is distinctly larger than andesine, which is 0.25 to 0.2 millimeters. Accessory minerals are quartz, magnetite, and sphene. Modes[1] are given in Table 1.

[1] All modes are estimates and are in volume per cent. Those calculated from chemical analyses are in weight per cent.

The plagioclase, many grains of which are slightly zoned, commonly shows albite twinning and, to a lesser extent, Carlsbad twinning. The composition ranges from intermediate andesine (An_{40}) to andesine-labradorite (An_{50}).[2] Optical data indicate that the amphibole is common hornblende, similar to that analyzed from an intrusive amphibolite in the Littleton-Moosilauke area (Table 10, column 18). A chemical analysis of amphibolite from the Mt. Washington area is given in Table 10, column 11.

Amphibolite, sufficiently rich in quartz, is classed as quartz amphibolite. The texture is granoblastic. These rocks (Table 1, column 2) differ from the more common amphibolites in containing orthoclase and pyroxene as well as quartz. Andsesine (An_{40}), is mostly untwinned although some shows albite twinning. The relative proportions of quartz and feldspar were determined by Dodge's (1936) method. The amphibole is common hornblende.

Biotite gneiss.—Biotite gneisses are the most abundant rocks in the Ammonoosuc volcanics of the Mt. Washington area. Megascopically they are gray and fine-grained, with variable amounts of biotite. In general they are even-grained, but some specimens show grains of feldspar and quartz several millimeters across set in a finer groundmass.

Microscopically the biotite gneisses consist essentially of quartz, plagioclase, potash feldspar, and biotite, with such minor accessories as muscovite, hornblende, apatite, garnet, and epidote. Modes are given in Table 1. The plagioclase is oligoclase and andesine, in which the anorthite content is constant in each specimen but ranges from 12 to 40 per cent in different specimens. Potash feldspar and plagioclase are present in all proportions.

Optical data indicate that most of the biotite is nearer annite than phlogopite, the weight per cent of iron equaling or exceeding the weight per cent of magnesia. The epidote, although not abundant, apparently is a primary metamorphic mineral. It occurs in small fresh anhedra, with about 13 per cent $HCa_2Fe_3Si_3O_{13}$, that is, 8 per cent Fe_2O_3. Muscovite is essentially $H_4K_2Al_6Si_6O_{24}$, similar to the analysis in Table 10, column 19.

By analogy with the Ammonoosuc volcanics in the low-grade metamorphic zone of the Littleton-Moosilauke area (Billings, 1937, p. 477), the biotite gneisses are considered to have been derived from pyroclastics. Many specimens show no trace of volcanic textures, but subhedral grains of feldspar and quartz, similar to those found in crystal tuffs, have been noted. No textures suggesting volcanic breccias or conglomerates were observed, but experience in the Littleton-Moosilauke area indicates that with increasing metamorphism such features tend to disappear. Assuming no important change in composition during metamorphism, rhyolites (Table 1, columns 3, 3a), quartz latites (column 4), and dacites (column

[2] In the laboratory study critical optical data were obtained for the principal minerals in every specimen for which a thin section was available; the data included the indices as determined by the immersion method. In a preliminary manuscript these data were recorded but have been omitted from the published manuscript. However, they are summarized graphically in Plate 8.

TABLE 1.—*Approximate modes—Ammonoosuc volcanics*

	1	1a	2	3	3a	4	5	6	7	8	9	10	11	12	13	14	15
Number of thin sections	2	..	1	2	..	1	2	1	1	2	1	2	1	2	1	1	1
Quartz	3	6	21	33	37	40	37	40	10	4	35	36	..	71	50	10	45
Orthoclase*	..	..	5	45	46	20	3	..	..	..	25	31	10	..	..	20	..
Plagioclase	27	37	43	10	10	33	44	40	58	45	32	23	17	5	32	35	12
Hornblende	68	55	25	..	..	1	..	..	8	tr	2	6	52	..	..	..	..
Pyroxene	..	..	5	..	..	..	..	..	20	27	1	..	..	..	..	..	..
Muscovite	..	..	..	1	2	..	3	2	..	..	..	..	..	20	3	..	15
Biotite	..	..	..	11	5	6	13	..	..	..	..	2	..	..	..	35	25
Garnet	..	..	..	tr	..	..	..	3	..	1	1	1	..	tr	..	..	3
Apatite	..	tr	tr	tr	tr	tr	tr	tr	..	tr	..	..	..	..	..	..	..
Epidote	..	..	..	..	..	tr	tr	..	3	22	4	tr	..	..	..	..	..
Chlorite	..	..	..	..	..	..	..	15	..	..	..	tr	18	..	15	..	..
Magnetite	1	2	..	..	..	..	..	..	..	..	..	..	1	..	..	..	..
Sphene	1	..	1	..	..	tr	..	..	1	1	..	1	1	..	tr	..	..
Pyrite	..	..	..	..	tr	..	..	..	..	..	..	..	..	4	..	..	..
Carbonate	..	tr	..	..	..	..	..	..	..	..	..	..	1	..	..	..	..
Sillimanite	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	tr
Per cent of anorthite in plagioclase																	
Average	45	52	40	12	10	35	32	13	40	45	35	57	46	5	38	30	15
Range	40–50	52	40±	12–13	10	35	25–40	13	40	40–50	35	35–90	46	5	38	30	15
Grain size millimeters	0.05–1.00	0.05–1.00	0.10–0.30	0.10–0.50	0.10–0.80	0.05–0.50	0.05–0.80	0.05–0.08	0.05–2.00	0.05–0.40	0.05–0.20	0.05–1.00	0.05–0.40	0.05–1.50	0.10–0.30	0.10–0.60	0.05–1.00

* Includes microcline and microperthite

1. Amphibolite.
1a. Amphibolite, calculated from chemical analysis (W 56).
2. Quartz amphibolite.
3. Biotite gneiss, originally rhyolite.
3a. Biotite gneiss, originally rhyolite, calculated from chemical analysis (W 26).
4. Biotite gneiss, originally quartz latite.
5. Biotite gneiss, originally dacite.
6. Chloritized biotite gneiss.
7. Pyroxene-hornblende granulite.
8. Pyroxene-epidote granulite.
9. Hornblende-epidote granulite.
10. Hornblende-biotite granulite.
11. Edenite-chlorite schist.
12. Quartz-muscovite-pyrite schist.
13. Quartz-chlorite-plagioclase gneiss.
14. Biotite-feldspar schist.
15. Mica schist.

5) are represented. Admixture of terrigenous material is suggested by the relatively high quartz or biotite content of some specimens.

A chemical analysis of biotite gneiss, which is metamorphosed rhyolite, is given in Table 10, column 14.

Lime-silicate granulite.—Granulites carrying hornblende and pyroxene, are not particularly abundant, but they are distinctive and have received special study. Moreover, megascopically they resemble lime-silicate granulites in the Fitch formation.

Megascopically the lime-silicate rocks are white to gray in weathered outcrops, gray to green in fresh exposures. Generally they are hard, resistant, and lack a good schistosity. Alternating bands, 1 to 10 millimeters thick differ in mineral composition. Pink garnets, black hornblende, and black pyroxene may be commonly recognized.

In the field these lime-silicate rocks are distinguished from the biotite gneisses by a lack of schistosity and the presence of hornblende and pyroxene. Their lighter color distinguishes them from the amphibolites. Some specimens closely resemble lime-silicate rocks of the Fitch formation. The means of distinguishing rocks from the two formations is discussed later.

Microscopically the lime-silicate rocks are granoblastic, although the hornblende commonly encloses the other minerals poikiloblastically. The individual minerals range considerably in size, but the pyroxene and hornblende are commonly 0.1 to 2.0 millimeters long, whereas the quartz and plagioclase are 0.05 to 0.2 millimeter. Modes are given in Table 1.

The pyroxene is a member of the diopside-hedenbergite series, and the content of diopside varies from 35 to 45 per cent. In most specimens the amphibole is common hornblende similar to that found in the Littleton-Moosilauke area. One specimen (Table 1, column 11) containing an amphibole with optical properties similar to edenite (Larsen and Berman, 1934, Table 6, No. 56), has obviously undergone hydrothermal alteration. Epidote, in subhedral to euhedral grains, appears to be contemporaneous with pyroxene, hornblende, and andesine. There is no evidence that it is a later hydrothermal alteration product. Optical data indicate 10 to 15 per cent of $HCa_2Fe_3Si_3O_{13}$.

The plagioclase is generally andesine, ranging from An_{35} to An_{50}, but in one specimen it is An_{90}. Chlorite from the edenite-chlorite schist is a later hydrothermal derivative from biotite, and its optical properties indicate that it is prochlorite (Winchell, 1933, p. 280).

The pyroxene-epidote granulites and probably the pyroxene-hornblende granulites were basaltic flows or tuffs. The edenite-chlorite schist, although considerably altered, also seems to have been originally a basalt. The hornblende-epidote and hornblende-biotite granulites, with their relatively high content of quartz and potash feldspar and low content of dark minerals, were presumably quartz latites.

The lime-silicate granulites of the Ammonoosuc volcanics superficially resemble the lime-silicate granulites of the Fitch formation. However, a distinction may be readily made microscopically. In the pyroxene of the Fitch formation γ is close to 1.708, whereas in the Ammonoosuc volcanics it averages 1.730. In the amphibole of the Fitch formation γ averages 1.648, whereas in the Ammonoosuc volcanics it is higher, ranging from 1.660 to 1.700. These differences are shown in Plate 8.

Quartz-muscovite-pyrite schist.—Although these rocks are not abundant, they are conspicuous and distinctive. In natural exposures they are rusty schists, stained deep brown by the alteration of pyrite to limonite. Fresh specimens are pure white, lustrous schists dotted with pyrite grains. Muscovite flakes are as much as 2 millimeters in diameter.

Under the microscope they are lepidoblastic to granoblastic, the grain size ranging from 0.05 to 1.50 millimeters. Quartz, muscovite, and pyrite are most conspicuous, with albite (An_5), garnet, and an unknown mineral as accessories. Modes are given in Table 1. The muscovite does not differ greatly from that found in the other formations.

These rocks apparently represent argillaceous sandstones interbedded with the volcanics. The pyrite was probably introduced hydrothermally during the metamorphism.

Mica schist and other schists.—Some of the Ammonoosuc volcanics are of sedimentary origin or represent mixtures of terrigenous and volcanic material. A coarse mica schist, containing small quantities of sillimanite (Table 1, column 15), was collected in the Moose River at an altitude of 1625 feet. The biotite is intermediate between phlogopite and annite. The muscovite is essentially $H_4K_2Al_6Si_6O_{24}$. This specimen carries sillimanite, indicating that the Ammonoosuc volcanics of the Mt. Washington area belong to the high-grade zone of metamorphism rather than to the middle-grade zone.

In a volcanic series composed dominantly of water-laid tuffs and breccias (Billings, 1937, p. 476) some admixture of terrigenous material is to be expected. The biotite-feldspar schist (Table 1, column 14), because of its high biotite content, is considered to represent a quartz latite tuff contaminated by argillaceous material. The quartz-chlorite-plagioclase gneiss (Table 1, column 13), is believed to represent a dacite contaminated by terrigenous quartz and argillaceous material. Furthermore, this specimen has undergone retrograde metamorphism. Some of the lime-silicate granulites and some of the biotite gneisses may be volcanics contaminated by sedimentary quartz.

RETROGRADE PROCESSES

In some specimens retrograde metamorphism is manifested primarily by the conversion of biotite to chlorite. Three specimens found within 100

FIGURE 1. TIMBERLINE
Looking north across mouth of Tuckerman Ravine toward Huntington Ravine. Shows transition from heavy timber through scrub timber into treeless zone.

FIGURE 2. ABOVE TIMBERLINE
Looking south toward Mt. Monroe from the south slope of Mt. Washington. Lakes of the Clouds in the right foreground. The Fitch formation occupies the depression at the foot of Mt. Monroe.

FIGURE 1. PARTRIDGE FORMATION
Alternating light and dark bands typical of the variety of gneiss which has been derived from shale without much change in chemical composition.

FIGURE 2. LITTLETON FORMATION
Coarse rough pseudo-andalusite schist. Large pseudo-andalusite crystals, which now consist of sillimanite, muscovite, and quartz, with a little staurolite.

FIGURE 3. LITTLETON FORMATION
Interbedded quartzite and spangled muscovite schist.

METAMORPHIC ROCKS

feet of each other at an altitude of 1550 to 1560 feet on Moose River, contain considerable prochlorite or ripidolite. One of these specimen is also rich in edenite. The three specimens lie within the Pine Mountain fault zone. Presumably fractures permitted the access of hydrothermal solutions necessary for these retrograde changes.

THICKNESS

The Ammonoosuc volcanics in the Mt. Washington area attain a maximum thickness of 5000 feet; the base, however, is not exposed because of the intrusive relations of the Oliverian magma series. A fairly complete section along Israel River shows dips almost exclusively to the southeast, although some northwesterly dips may be observed. As shown in section CC′ of Plate 11, the thickness approximates 5000 feet. Along the headwaters of Moose River the thickness is likewise 5000 feet. On other streams the range is from 3500 to 5000 feet.

A rough estimate of the relative abundance of the various types is as follows: amphibolite 3 per cent; biotite gneiss 93 per cent; lime-silicate granulites 1 per cent; quartz-muscovite-pyrite schist 3 per cent; mica schist, trace.

PARTRIDGE FORMATION (ORDOVICIAN?)

GENERAL STATEMENT

The Partridge formation overlies the Ammonoosuc volcanics and occupies two areas (Pl. 1), one of which extends northeasterly from Mt. Deception through Mt. Bowman to the northeast corner of the map. The second area, forming a great irregular semicircle, may be traced along the east side of the map from Dolly Copp Camp to Pinkham Notch, thence westward to Mt. Franklin and northwesterly to Millen Hill.

Good exposures may be seen in many places, especially at: (1) Peabody River, at and below the junction with the West Branch; (2) West Branch of the Peabody River, for 1 mile above the junction with the main stream; (3) Cutler River, just above Pinkham Notch; (4) summit of Mt. Clay; (5) summit of Mt. Monroe.

LITHOLOGY

General character.—The Partridge formation consists of two major types: paraschists and gneisses. The paraschists are quartzites and mica schists, derived respectively from sandstones and shales. The gneisses, which are typically composed of alternating dark and light bands, have also been derived from shales. Closely associated with the gneisses are small bodies of granitoid rocks. The gneisses are far more common than the paraschists. However, in places, notably along the lower part of the West Branch of the Peabody River and the Cutler River just west of

Pinkham Notch, the formation is paraschist rather than gneiss. Because of the scale of the map (Pl. 1) not all the details of distribution of the lithologic types could be illustrated.

Gneiss.—The gneiss is composed of alternating light and dark layers (Pl. 5, fig. 1). The dark layers consist essentially of biotite, muscovite, quartz, and accessories, whereas the light layers are primarily plagioclase and quartz, with only minor amounts of biotite and muscovite. Individual layers range in thickness from 0.5 to 3 centimeters. The ratio of dark to light material differs greatly. Wherever the lighter material comprises only a small percentage of the whole rock, the plagioclase-quartz aggregate occurs as short pods which in cross section appear to be from 1 to 10 centimeters long. Where the light and dark constituents are equal in amount, both are discontinuous, mutually interfingering layers. With a still greater increase in the lighter constituents, the dark layers are short and discontinuous. In the North Conway quadrangle (Fig. 1) large areas of such rocks are mapped as part of Chatham granite (Billings, 1928, p. 82-83). In the present paper they are mapped as the gneissic portion of the Partridge formation.

The individual minerals range in diameter from a fraction of a millimeter to 5 millimeters. The muscovite flakes are the largest and most conspicuous. The dark, biotite-rich layers are schistose parallel to the gneissic banding. The light-colored, plagioclase-quartz aggregates, however, are massive and typically granitoid. Two lines of evidence indicate that the banding is parallel to the original bedding. First, the banding is parallel to the bedding in the underlying Ammonoosuc volcanics and overlying paraschists of the Partridge, Fitch, and Littleton formations. Second, the bedding in small patches of paraschist preserved within the gneiss is parallel to the gneissic banding.

Locally, the gneissic banding is thrown into folds which are similar to those in the paraschists of the Partridge and Littleton formation. The folding must have preceded the final recrystallization, as the individual minerals are not granulated and the light-colored layers are typically granitoid.

The mineralogical composition of the gneiss is summarized in Table 2. Column 1 gives its composition as a whole; column 2 represents the dark portion; and column 3 gives the light portion.

In those specimens in which the light constituents are in isolated lenses, the dark portions are black to gray mica schists that do not differ greatly from some of the paraschists. Biotite, muscovite, quartz, garnet, and sillimanite are important. More commonly, where the light and dark bands are approximately equal, although the alternating layers are clearly defined in the hand specimens, the dark layers contain considerable plagioclase.

The light layers consist largely of plagioclase and quartz with minor amounts of biotite and muscovite. The texture is essentially hypidiomorphic granular. Within the limits of a thin section the segregation into dark and light layers may not be apparent.

The anorthite content of plagioclase ranges from 15 to 30 per cent in different specimens. In all the specimens of gneiss studied microscopically, not one grain of orthoclase or microcline has been observed. Muscovite is similar to those obtained in other formations. The biotite is not greatly different from that in the paraschists of the Partridge and Littleton formations.

A chemical analysis of gneiss in which the dark and light layers are about equal is given in Table 10, column 10.

Granitoid rocks.—Patches of granitic rocks too small to be shown on Plate 1 occur in the gneiss at the following localities: (1) Peabody River, between altitudes of 1700 and 1800 feet; (2) one mile southwest of Pinkham Notch, on New River, between altitudes of 2780 and 2970 feet and between 3270 and 3370 feet.

They are massive, homogeneous, gray rocks composed of plagioclase, quartz, biotite, and muscovite, with conspicuous garnet in some specimens. The characteristic grain-size is 1 to 3 millimeters but in the exposures on the Peabody River large muscovite flakes, up to 1 centimeter across, flash conspicuously in the sunshine. The muscovite encloses other minerals poikiloblastically. These granitoid rocks resemble the nonporphyritic phase of the Kinsman quartz monzonite of the Littleton-Moosilauke area (Billings, 1937, p. 506-507).

Three modes are given in Table 2, columns 4, 5, and 6. One is quartz diorite, a second is granodiorite, and a third is quartz monzonite. In naming these rocks the Johannsen system has been followed, except that muscovite has been treated as if it were potash feldspar. The average mineral composition is not greatly different from that of the gneiss.

That these granitoid rocks are younger than the gneiss of the Partridge formation may be demonstrated southwest of Pinkham Notch. At an altitude of 2850 feet on New River, angular inclusions of gneiss occur in a granitoid rock.

Paraschists.—The original sediments of the Partridge formation have not all been converted to gneiss. West of Pinkham Notch, along the lower part of the West Branch of Peabody River, and northeast of Boott Spur, the upper part of the Partridge formation consists of paraschists. Moreover, within the areas mapped as gneiss are small bodies of paraschist, too small to show on the map (Pl. 1).

The paraschists are lithologically identical with many of the types found in the Littleton formation and include: (1) shiny fine-grained

TABLE 2.—*Approximate modes—Partridge formation and associated plutonic rocks*

Type	1	1a	2	3	4	5	6	7	8	8a	9	10	11	12	13	14
Number of thin sections	10	..	5	5	1	1	1	1	2	..	1	2	2	2	1	1
Quartz	37	36	33	40	35	30	47	12	42	36	35	56	52	88	73	54
Plagioclase	36	17	17	47	49	45	18	..	3	15	..	5	5	6	..	32
Biotite	15	29	27	5	10	15	6	20	14	29	..	8	5	5	7	..
Muscovite	11	12	19	7	3	10	20	63	29	9	10	18	27	1	12	..
Garnet	1	1	3	tr	..	tr	1	..	..	1	1	1	..	tr	tr	5
Sillimanite	tr	5	1	1	..	..	tr	5	..	10	3	1	..	..	..	..
Magnetite	tr	..	tr	tr	..	..	tr	tr	..	..	3	1	..	tr	..	..
Apatite	tr	tr	tr	tr	..	tr	..	tr	..	tr	..	tr	tr	tr	..	..
Staurolite	..	..	..	..	..	..	..	tr	..	..	..	..	..	..	tr	..
Tourmaline	..	..	..	tr	..	..	..	tr	tr	..	..	..	tr	tr	tr	..
Chlorite	tr	..	tr	tr	..	..	8	..	3	..	16	10	2	tr	8	..
Pyrite	tr	tr	..	..	3	tr	..	..	4	tr	..	..	8	tr	..	..
Sericite	tr	..	..	..	..	..	..	..	5	..	32	..	1	tr	..	..
Sphene	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	1
Zircon	tr	..	tr?	..	..	..	tr	..	..	..	..	..	..	tr	tr	..
Hornblende	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	8
Per cent of anorthite in plagioclase																
Average	19	19	22	20	21	26	15	..	n.d.	24	..	16	40	15	..	92
Range	15–30	19	18–30	16–30	21	26	15	..	n.d.	24	..	16	40	15	..	92
Grain size in millimeters	0.03–6.00	0.03–4.00	0.10–6.00	0.10–6.00	0.30–4.00	0.20–0.50	0.50–6.00	0.50–4.00	0.05–0.80	..	0.05–3.00	0.10–4.00	0.05–1.00	0.05–1.50	0.10–2.50	0.05–1.00

1. Gneiss.
1a. Gneiss, calculated from chemical analysis, specimen W215.
2. Dark portion of gneiss.
3. Light portion of gneiss.
4. Quartz diorite.
5. Granodiorite.
6. Quartz monzonite.
7. Muscovite—rich knot in a coarse rough pseudo-andalusite schist.
8. Shiny fine-grained pseudo-andalusite schist.
8a. Shiny fine-grained pseudo-andalusite schist, calculated from chemical analysis, specimen W229.
9. Medium-grained massive schist.
10. Spangled muscovite schist.
11. Shiny fine-grained mica schist.
12. Quartzite.
13. Quartz conglomerate.
14. Lime-silicate concretion.

pseudo-andalusite schist; (2) medium-grained massive schist; (3) spangled muscovite schist; (4) shiny fine-grained mica schist; (5) quartzite; and (6) a few thin beds of fine-grained quartz conglomerate. Modes of the various types are given in Table 2. A detailed description is unnecessary, as similar rocks are described under the Littleton formation. Because of variations in the conditions of metamorphism certain types found in the Littleton formation—notably those containing staurolite, large andalusite crystals, or pseudomorphs after large andalusite crystals —are not found in the paraschists of the Partridge formation. On Millen Hill, however, there is a small area underlain by coarse rough pseudo-andalusite schist, which is similar to rocks in the Littleton formation.

A chemical analysis of shiny fine-grained pseudo-andalusite schist is given in Table 10, column 4.

THICKNESS

The thickness of the Partridge formation can be determined most satisfactorily along the Israel River, northeast of Mt. Bowman and along some of the small streams for a mile to the northeast. This is the only place in the whole area where the Partridge formation is in contact with the overlying and underlying stratigraphic units. The formation here consists entirely of gneiss. The structure is comparatively simple, for the foliation of the gneiss, as well as the bedding of the overlying and underlying formations, dips uniformly southeast. The thickness of the gneiss is approximately 1400 feet, but the thickness of the original shale from which the gneiss has been derived is conjectural. Considerable tectonic thinning may have occurred; on the other hand, there may have been some thickening by the introduction of magmatic material.

FITCH FORMATION (SILURIAN)

GENERAL STATEMENT

The Silurian Fitch formation, although comparatively thin, is a great aid in solving the geological structure of the Presidential Range. This unit is best developed south of Mt. Washington and on the west side of the Peabody River valley between Pinkham Notch and Dolly Copp Camp. Unusually good and accessible localities to see the formation are: (1) West Branch of the Peabody River, 1620 to 1635 feet; (2) Cutler River, 2170 feet, between the Appalachian Mountain Club bridge and the foot of Crystal Cascade; (3) pass 1000 feet northeast of summit of Mt. Monroe; and (4) headwall of Oakes Gulf. In general the Fitch formation is 20 to 50 feet thick. Locally it is 200 feet thick but is absent at the appropriate horizon on the northern slopes of Mount Adams and Mount Madison.

14. Lime-silicate concretion.

7. Muscovite—rich knob in a coarse rough pseudo-andalusite schist.
8. Shiny fine-grained pseudo-andalusite schist.

LITHOLOGY

General character.—The rocks of the Fitch formation consist primarily of diopside, actinolite, biotite, plagioclase, microcline, and quartz, with such accessories as hornblende, sphene, clinozoisite, apatite, garnet, pyrite, zircon, muscovite, and tourmaline. On the basis of equilibrium relations the rocks may be classified as diopside granulite, diopside-actinolite granulite, actinolite granulite, actinolite-biotite granulite, biotite schist, and biotite-pyrite schist. Modes of rocks microscopically studied are in Table 3. In general the various types are in relatively thin beds, a fraction of an inch to 3 inches thick. Locally, however, the beds may be 3 feet thick, and some of the biotite schists are 10 feet thick.

The granulites, because of the diopside and actinolite, are believed to have been derived from dolomitic sandstones and shales (Billings, 1937, p. 486, Table 6). The schists are believed to represent calcareous shales.

Diopside granulite.—The diopside granulites weather pure white and on hasty inspection resemble quartzite. Fresh specimens are white with a slight green tinge and have a fine-grained, granular texture, the individual grains averaging a fraction of a millimeter to 2 millimeters in diameter. The dominant and diagnostic mineral is diopside, whereas plagioclase and microcline are of variable importance. Accessories include sphene, actinolite, and biotite. Modes are given in Table 3.

Diopside-actinolite granulite.—Eight of the rocks studied microscopically are diopside-actinolite granulites. In some of these specimens green needles of actinolite, 1 to 3 millimeters long, lie in a greenish-gray groundmass. The needles tend to parallel the bedding and the long axes are aligned, giving the rock a distinct lineation. In other specimens, more uniformly greenish-gray, the actinolite needles are less conspicuous as they are only a millimeter long, but inspection with the hand lens shows them clearly. In these specimens the needles parallel the bedding, but in some they are diversely oriented within these planes, in others the long axes are aligned.

Microscopic study shows that in addition to actinolite and diopside, the essential minerals are microcline, plagioclase, and quartz, with such accessories as biotite, sphene, and clinozoisite. Modes are given in Table 3. The texture is granoblastic, except for the poikiloblastically developed actinolite which is in larger needles (1 to 4 millimeters) than the other minerals (0.1 to 1.0 millimeter).

Actinolite granulite.—The actinolite granulites do not differ megascopically from the diopside-actinolite granulites. In some, conspicuous dark-green actinolite needles, 1 to 3 millimeters long, are set in a greenish-gray groundmass. In others, the rock is a uniform greenish-gray, and a hand lens is necessary to discern actinolite needles a millimeter long.

Microscopic studies show the important constituents to be plagioclase, microcline, and quartz with such accessories as sphene, clinozoisite, biotite, and apatite. The texture is essentially granoblastic but the actinolite needles surround the other minerals poikiloblastically.

TABLE 3.—*Approximate modes—Fitch formation*

	1	2	3	4	5	6
Number of thin sections averaged	2	8	4	3	7	1
Quartz	....	11	21	22	27	40
Microcline	9	22	16	....	13	3
Plagioclase	19	10	31	13	25	10
Diopside	69	29	....	....	....	....
Actinolite	tr	23	29	34	....	....
Biotite	1	4	1	26	32	32
Sphene	2	1	2	2	tr	tr
Garnet	....	....	tr	1	1	5
Clinozoisite	....	tr	tr	....	....	....
Apatite	....	....	tr	tr	tr	tr
Pyrite	tr	....	tr	2	1	10
Zircon	....	....	tr	....	....	....
Magnetite	....	....	....	tr?	....	....
Muscovite	....	....	....	....	1	....
Tourmaline	....	....	....	....	tr	....
Per cent of anorthite in plagioclase:						
Average	70	82	53	77	34	92
Range	55–86	55–92	20–86	77	28–47	....
Grain size in millimeters	0.05–2.00	0.05–4.00	0.05–1.50	0.05–1.00	0.05–2.00	0.05–0.10

(1) Diopside granulite.
(2) Diopside-actinolite granulite.
(3) Actinolite granulite.
(4) Actinolite-biotite granulite.
(5) Biotite schist.
(6) Biotite-pyrite schist.

Actinolite-biotite granulite.—The actinolite-biotite granulite is a brownish or purplish rock, interbedded with thin layers of actinolite granulite. The elliptical biotite flakes have parallel long axes, producing a distinct lineation. The rock is granoblastic; its mode is shown in Table 3.

Biotite schist.—The biotite schists are black to brownish-gray schistose rocks, usually with a well-marked lineation due to the parallel long axes of elliptical biotite plates.

Microscopic study confirms the lepidoblastic texture. Biotite and quartz are essential constituents; plagioclase is present in varying amounts. Microcline occurs in some specimens. Accessories include muscovite, garnet, and tourmaline. Modes are given in Table 3.

Biotite-pyrite schist.—The biotite-pyrite schists are conspicuous in the field because of their rusty brown color, due to limonitization of the pyrite. Fresh surfaces are gray, and the rock has a fair schistosity on which lineation is evident, due to the elliptical flakes of biotite. A mode is given in Table 3.

TABLE 4.—*Thickness of the Fitch formation*

Locality	Measured thickness (in feet)	Probable thickness (in feet)
West Branch of Peabody River, 1620-1635 feet	136	200?
1600 feet on stream, 2800 feet NW. of Glen House	25	50?
Emerald Pool, ½ mile SW. of Glen House	27	27
Peabody River, 2380 feet.....	15±	15±
Cutler River, 2200 feet, just below Crystal Cascade	65	65
Headwall of southern amphitheatre of Tuckerman Ravine	10 5/6	12+
500 feet SW. of Boott Spur	Present only in the float	100?
Headwall of Oakes Gulf	38.5	50?
4445 feet on east branch of stream flowing north from BM 5305 on Mt. Washington Auto Road	28.5±	28.5+
Ravine of the Castles, ½ mile east of Mt. Bowman	30±	30±

MINERALOGY

Diopside is remarkably uniform in the 10 specimens in which it was found. Optical data indicate from 16 to 22 per cent of hedenbergite in solid solution (Winchell, 1933, p. 226). Actinolite is likewise remarkably uniform in the 15 specimens in which it was found. Optical data indicate from 18 to 25 per cent of $CaFe_5Si_4O_{12}F$ in solid solution (Winchell, 1933, p. 246). Biotite of the Fitch formation is more variable in composition than the actinolite and diopside. In those rocks carrying diopside and actinolite, however, the indices are systematically lower and the pleochroic colors less intense than in rocks devoid of those minerals. In rocks with actinolite and diopside the biotite averages 72 per cent phlogopite, and the range is 69 to 78 per cent (Winchell, 1933, p. 274). In specimens without lime silicates, the biotite is apparently richer in iron, and the data suggest an average phlogopite content of 59 per cent, ranging from 50 to 67 per cent. Although the indices of biotite

TABLE 5.—*Measured Sections of the Fitch formation*

(1) WEST BRANCH OF PEABODY RIVER

800 feet downstream from the Osgood bridge, altitude 1620-1635 feet. (Stratigraphically highest units, which are at the southwest end of the outcrop, are at top of table)

	Thickness (feet)
Green actinolite granulite *	17
Black biotite schist	6
Green actinolite granulite *	3
Biotite schist	8
Green actinolite granulite *	2
Dark biotite schist	6
Green actinolite granulite *	5
Dark biotite schist	3
Green actinolite granulite *	½
Biotite schist	9
Biotite schist, some beds containing small white needles 2 to 6 millimeters long	13
Interbedded white diopside granulite and green actinolite granulite *, the individual beds about 1 inch thick	15
Dark biotite schist	1
Interbedded white diopside granulite and green actinolite granulite*	½
Biotite schist	22
Interbedded white diopside granulite and green actinolite granulite*	5
Biotite schist	5
Actinolite granulite *	3
Interbedded biotite schist and actinolite granulite *	12
Total	136

(2) CUTLER RIVER

2200 feet, just below Crystal Cascade.

		Thickness (feet)
	Mica schist above Fitch formation	
Fitch formation	Lime-silicate granulites	20
	Interbedded lime-silicate granulites and biotite schists	40
	Rusty biotite-pyrite schist	5
	Granite and pegmatite	
	Total	65

(3) TUCKERMAN RAVINE

400 feet southwest of where Tuckerman Ravine Trail crosses the "Little Headwall"; ½ mile northeast of Boott Spur.

		Thickness (feet)
	Spangled muscovite schist above Fitch formation	
Fitch formation	Biotite schist	0.5
	Thin-bedded actinolite-biotite granulite	4
	Gray biotite gneiss	1
	Thin-bedded actinolite granulite	1
	Gray mica schist	1.2
	Pegmatite sill	3.5
	Biotite schist	0.1
	Pegmatite sill	6
	Actinolite-diopside granulite	3
	Base not exposed.	
	Total	10.8†

* Rocks classified as actinolite granulite in the field undoubtedly contain diopside, but inasmuch as no microscopic study was made of these particular rocks, definite data are unavailable. The actinolite granulites of this table correspond therefore to the diopside-actinolite and actinolite granulites of Table 3.

† Pegmatites are not included in totals.

TABLE 5.—*Measured Sections of the Fitch formation—Continued*

(4) ALTITUDE OF 4750 FEET

Headwall of Oakes Gulf, 2800 feet southeast from the Lakes of the Clouds.

		Thickness (feet)
	Quartzite and muscovite schist above Fitch formation	..
Fitch formation	Biotite schist	5
	Platy diopside-actinolite granulite and some biotite schist	10
	Fine-grained biotite schist	1.5
	Biotite schist, somewhat injected	2
	Actinolite-diopside granulite	7
	Gap	5
	Biotite schist	3
	Pegmatite	0.5
	Actinolite-diopside granulite	2.5
	Gray biotite schist	1.5
	Dark biotite schist	1
	Base not exposed.	
	Total	38.5 †

(5) ALTITUDE 4445 FEET

On east branch of stream flowing north from BM 5305 on Mount Washington Auto Road.

		Thickness (feet)
	Gneiss in Littleton formation	..
Fitch formation	Actinolite granulite	0.9
	Gneiss	17.3
	Actinolite granulite	5.5
	Interbedded quartzite and actinolite granulite	23.0
	Quartzite in Partridge formation	
	Total	46.7

† Pegmatites are not included in totals.

are not very reliable in determining chemical composition, they indicate the trend in composition.

Plagioclase, as noted in Table 3, differs in composition, the anorthite content ranging from 20 to 92 per cent. The amount in the plagioclase is independent of the other minerals present. Muscovite is found in only two of the specimens, and the optical data indicate essentially pure $H_4K_2Al_6Si_6O_{24}$.

THICKNESS AND MEASURED SECTIONS

The thickness of the Fitch formation was determined in a number of localities. In some of these either the top or the bottom of the formation is not exposed so that the measured thickness is a minimum value. In Table 4 the measured thickness is given in one column, the probable thickness in a second column. In places, notably on the northern slopes of Mt. Adams and Mt. Madison, the Fitch formation is absent at the appropriate horizon. The maximum measured thickness is 136 feet, and the maximum probable thickness is 200 feet. The arithmetical mean of the "probable thickness" is 60 feet. Several stratigraphic sections were measured and the data are given in Table 5.

LITTLETON FORMATION (DEVONIAN)

GENERAL STATEMENT

The Littleton formation, the highest stratigraphic unit in the area, occupies a large portion of the eastern part of the geological map (Pl. 1) and is the bedrock on most of the higher summits of the Presidential Range. Only locally, as in the vicinity of Mt. Clay, are other formations found.

Excellent exposures may be seen: (1) along the Mt. Washington Auto Road; (2) Mt. Madison; (3) ridge south of Cutler River; (4) Tuckerman Ravine; (5) Huntington Ravine; (6) half a mile northeast of Mt. Monroe; (7) pass south of Mt. Jefferson; (8) pass north of Mt. Jefferson; (9) ridge 1½ miles southwest of Dolly Copp Camp.

The Littleton formation consists largely of paraschists; gneiss occurs locally in the lower part of the formation. Many of the schists and gneisses are so similar to those in the Partridge formation that a distinction can be made only by stratigraphic relations—that is, whether the beds lie above or below the Fitch formation.

PARASCHISTS

General statement.—The paraschists of the Littleton formation consist of schists and quartzites. In many localities bedding is conspicuous, due to alternating layers of different types (Pl. 5, fig. 3). The thickness of the strata ranges from an inch or less up to 10 feet. The schistosity characteristically parallels the bedding, but locally fracture cleavage at an angle to the bedding is prominent.

The argillaceous types, originally shale, are varied primarily because of differences in the metamorphic conditions and secondarily because of differences in the original rock composition. They may be most conveniently classified as: (1) coarse andalusite schist; (2) coarse pseudo-andalusite schist; (3) coarse rough pseudo-andalusite schist; (4) intermediate pseudo-andalusite schist; (5) shiny fine-grained pseudo-andalusite schist; (6) staurolite schist; and (7) medium-grained massive schist. Less aluminous shales are now (8) spangled muscovite schist, and (9) shiny fine-grained mica schist. The original arenaceous shales or argillaceous sandstones are now (10) mica-quartz schist. The original sandstones are (11) quartzites; (12) pink garnet granulites, although thin, are conspicuous.

Coarse andalusite schist.—The coarse andalusite schists and coarse pseudo-andalusite schists occupy several square miles northwest of Pinkham Notch (Fig. 3). The least altered andalusite schists have large, unoriented porphyroblasts of andalusite, 1 to 8 centimeters long and a half

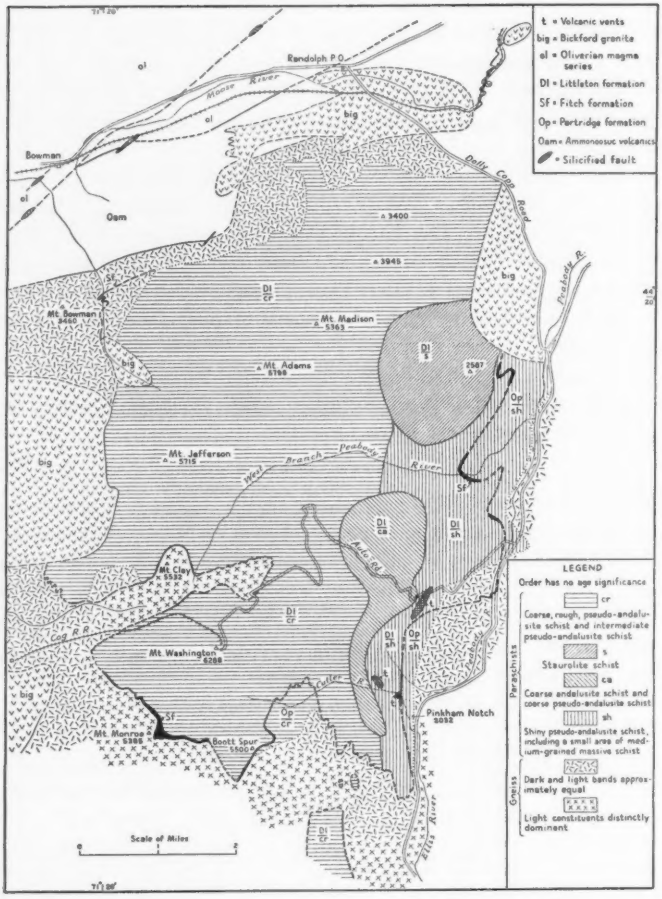


FIGURE 3.—*Metamorphic subzones in Mt. Washington area*

Sediments that were originally the same kind of shale now vary greatly in appearance because of differences in mineralogy or texture, inasmuch as the character of the metamorphism was not the same everywhere. The legend in the lower right-hand corner, by pattern and letter, shows the present character of the original shale. The legend in the upper right-hand corner shows, by letters, the formation to which the rocks belong. In the letter-symbols on the map the letters above the line indicate the formation, the letters below indicate the present character of original shales.

to 1 centimeter thick, set in a fine-grained shiny matrix of mica schist (Pl. 6, fig. 1). Andalusite has well-developed prismatic cleavage and is flesh-colored except for a darker central core. Even in the least altered specimens, an outer shell has been converted to muscovite, with some associated staurolite. Some specimens show sillimanite closely associated with andalusite, particularly as an outer shell. The matrix is shiny mica schist composed of muscovite, biotite, quartz, and garnet. A mode of the coarse andalusite schist appears in Table 6, column 1.

Coarse pseudo-andalusite schist.—The coarse pseudo-andalusite schist differs megascopically from the coarse andalusite schist in the gray, greasy lustrous color of the large columnar pseudo-andalusite crystals. Microscopic study shows the andalusite to be completely sericitized. The groundmass is lustrous mica schist, strongly crumpled in many localities.

Table 6, column 2 represents the mineralogy of these rocks. In one specimen the muscovite is in two generations. Many of the larger flakes (0.5 to 1.0 millimeter across) and all the smaller ones (0.03 to 0.02 millimeter across) have replaced andalusite. Sillimanite is associated with the smaller muscovite flakes, but the exact amount is hard to determine because the birefringence and character of the elongation of sillimanite and muscovite are similar.

Coarse rough pseudo-andalusite schist.—The coarse rough pseudo-andalusite schists characterize the higher parts of the Presidential Range. Large knots, from 2 to 20 centimeters long and a half to 2 centimeters wide, are irregularly distributed in a schistose matrix (Pl. 5, fig. 2, Pl. 6, fig 2.). The knots are muscovite and sillimanite, with lesser amounts of staurolite and sericite. The individual muscovite flakes are 2 to 10 millimeters across, and the sillimanite crystals 5 to 20 millimeters long. Generally the sillimanite is in the interior of the knot, whereas the muscovite forms the outer shell. Staurolite, in crystals 1 to 2 millimeters long, is locally associated with the muscovite in the outer part of the knot but more commonly is irregularly distributed through the matrix of the rock. These "knots" are pseudomorphs after andalusite. Not only do they have the form of andalusite, but the description of the coarse andalusite schist, already given, makes it clear that all stages in the transition from andalusite to the knots may be observed.

The groundmass of the coarse rough pseudo-andalusite schist is typical mica schist, composed of muscovite, biotite, garnet, tourmaline, and quartz. Modes are given in Table 6, columns 3 and 3a; a chemical analysis appears in Table 10, column 5.

Intermediate pseudo-andalusite schist.—The intermediate pseudo-andalusite schists are extensively developed northeast of Mt. Monroe. The pseudo-andalusite crystals, which are 1 to 4 centimeters long and 2 to 5 millimeters wide (Pl. 6, fig. 3), are smaller than those in the coarse rough pseudo-andalusite schist.

Characteristically, but not necessarily, the long axes of the crystals parallel one another, giving the rock a distinct lineation parallel to the pitch of the fold axes. The pseudo-andalusite crystals are largely irregular patches of sillimanite associated with muscovite. Large irregular flakes of muscovite 5 to 10 millimeters across are distributed through the rock. The groundmass, essentially quartz, muscovite, and biotite, contains stubby crystals of black tourmaline 1 to 7 millimeters long. Column 4 of Table 6 illustrates this variety of schist.

Shiny fine-grained pseudo-andalusite schist.—With progressive diminution in the size of the pseudo-andalusite crystals the intermediate variety passes into the shiny variety. The shiny pseudo-andalusite schists are typically developed on the eastern slopes of the Presidential Range (Fig. 3); excellent exposures occur in the Cutler River, above Crystal Cascade, and on Peabody River, between altitudes of 2400 and 2900 feet.

These rocks are characterized by shiny planes of schistosity which glisten in the sunlight. Although the fresh surfaces are invariably gray and the weathered surfaces usually so, some beds are rusty brown due to oxidation of pyrite. Schistosity, parallel to the bedding, is generally well developed and in many localities has been thrown into a series of small crinkles or folds with a wave length of 3 to 10 millimeters and an amplitude of 1 to 2 millimeters.

Small white crystals of pseudo-andalusite, 5 to 10 millimeters long, 2 to 4 millimeters wide, and a half to 1 millimeter thick, lie with their greatest and intermediate axes in the plane of schistosity; in many instances the long axes are parallel and give the rock a distinct lineation. On fresh surfaces the pseudo-andalusite has a greasy lustre, and in sections perpendicular to the lineation it has the appearance of augen.

Inspection with the hand lens shows that the groundmass of the rock contains considerable biotite and muscovite 0.1 to 3 millimeters across. Subhedral garnets up to 0.5 millimeters are present in some specimens. Locally these schists are greenish-gray due to chloritization of biotite.

An average mode of the shiny fine-grained pseudo-andalusite schist is given in Table 6, column 5. The texture is lepidoblastic, with the pseudomorphs of andalusite in distinct augen. Microscopic study shows these crystals to be chiefly sericite with lesser amounts of sillimanite and muscovite. Unaltered andalusite has not been observed but that this was the original mineral is demonstrated by similar relations in the coarse pseudo-andalusite schists. Sillimanite, present in the pseudomorphs

FIGURE 1. COARSE ANDALUSITE SCHIST
Large clear crystals of pink andalusite in a groundmass of fine-grained mica schist. Scale in inches.

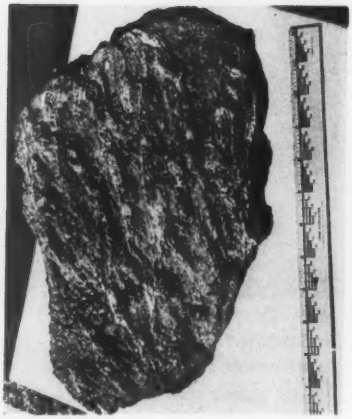


FIGURE 2. COARSE ROUGH PSEUDO-ANDALUSITE SCHIST
Shows large pseudo-andalusite crystals similar to Figure 2 of Plate 5.

FIGURE 3. INTERMEDIATE PSEUDO-ANDALUSITE SCHIST
Long, slender, pseudo-andalusite crystals, now sillimanite, muscovite, and some staurolite. Note excellent lineation.

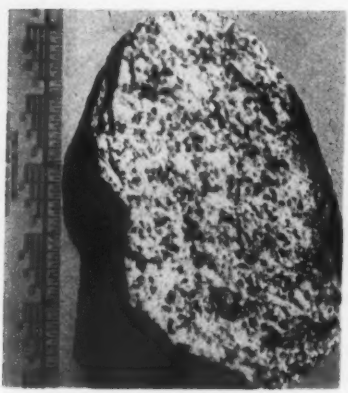


FIGURE 4. BICKFORD GRANITE
Shows typical hypidiomorphic granular texture.

LABORATORY SPECIMENS

FIGURE 1. SERIES OF MINOR FOLDS
Axial planes are nearly vertical and axes have gentle plunge.

FIGURE 2. PLUNGING FOLD
Axial plane dips away from reader and axis plunges 25 degrees to left.

FIGURE 3. CLOSE-UP OF FOLD
Axial plane is steep and plunge of axis is gentle.

FIGURE 4. COARSE ROUGH PSEUDO-ANDALUSITE SCHIST
This fold shows folded and bent pseudo-andalusite crystals.

FOLDS IN THE LITTLETON FORMATION

after andalusite, also occurs in small needles throughout the rock. Locally the pseudomorphs consist exclusively of sillimanite. Some specimens have undergone retrograde metamorphism, so that all the biotite is now chlorite.

Staurolite schist.—The staurolite schists are exposed on the lower eastern slopes of Mt. Madison (Fig. 3). The best exposures are on the ridge 1½ miles southwest of Dolly Copp Camp. Small euhedral to subhedral porphyroblasts of staurolite, 1 to 5 millimeters long, are set in a light-gray to dark-gray matrix. Cruciform twins, on (032) which gives a rectangular cross, and on (232) which gives a cross at 60 degrees, are present. Rarely the staurolite is surrounded by a thin shell of sericite 0.1 millimeter thick.

Some specimens contain andalusite or pseudomorphs after andalusite 1 to 10 centimeters long and 4 to 8 millimeters wide. Some crystals have a core of gray andalusite surrounded by a shell of flesh-colored andalusite, outside of which is a replacement shell composed of muscovite, staurolite, and grains of quartz 1 to 3 millimeters long. In many cases the andalusite has been completely replaced by this aggregate. The gray shiny groundmass of these schists is biotite, muscovite, and quartz, with some garnet. A mode is given in Table 6, column 6.

Medium-grained massive schists.—In a small area west of the Glen House, extending up to the Mt. Washington Auto Road to an altitude of 2400 feet (Fig. 3), the schists are massive rather than schistose and have an average grain size of several millimeters. The essential minerals are muscovite in conspicuous flakes, biotite, and quartz. Sillimanite, in stubby crystals 4 to 8 millimeters long, is prominent in some specimens and in places displays a distinct lineation. Retrograde processes are indicated by sericitization of the sillimanite, and chloritization of the biotite.

These massive schists do not differ chemically from the shiny fine-grained pseudo-andalusite schists. The differences, one schistose, the other massive, are due to tectonic reasons. The shiny fine-grained pseudo-andalusite schists were in an area of strong rock flowage, whereas the massive schists were not.

Spangled muscovite schist.—All of the paraschists of the Littleton formation described in the preceding pages were initially of the same chemical composition, inasmuch as they were aluminous shales. Aluminous minerals, such as andalusite, pseudo-andalusite, sillimanite, and staurolite, are conspicuous. Less aluminous shales have been metamorphosed to spangled muscovite schist and shiny mica schist. These rocks are not devoid of aluminous minerals, but such minerals are not common.

The spangled muscovite schists are best developed in areas occupied by the coarse rough pseudo-andalusite schists, the intermediate pseudo-andalusite schists, the coarse andalusite schists, and the coarse pseudo-andalusite schists.

Typically, the spangled muscovite schists have large porphyroblasts of muscovite, 2 to 10 millimeters in diameter, in a gray massive to schistose groundmass of biotite, muscovite, and quartz, with small amounts of garnet and tourmaline. All transitions to the coarse rough pseudo-andalusite schists exist, and hand specimens of the two rocks may not differ much in appearance. However, the spangled muscovite schist lacks the pseudomorphs after andalusite characteristic of the coarse rough pseudo-andalusite schist. A mode is given in Table 6, column 8.

Shiny fine-grained mica schists.—The shiny fine-grained mica schists occur in the same areas as the shiny fine-grained pseudo-andalusite schists and represent the less aluminous shales. Fresh surfaces are always gray, but weathered surfaces of many specimens are rusty brown due to oxidation of pyrite. The essential minerals are muscovite, biotite, and quartz. The schistosity, which parallels the bedding, displays a well-defined lineation, due to the alignment of the long axes of oval-shaped grains of biotite or to small crinkles or folds with a wave length of 2 to 3 millimeters and an amplitude of 0.2 to 0.5 millimeter. In some specimens muscovite porphyroblasts, 2 to 3 millimeters across, lie within the plane of schistosity, and the rock similates the spangled muscovite schist. A mode is given in Table 6, column 9.

Mica-quartz schist.—Following the usage employed in the Littleton-Moosilauke area (Billings, 1937, p. 472), rocks with 60 to 80 per cent quartz (or quartz plus feldspar, if the feldspar is not abundant) are termed mica-quartz schists and were derived from arenaceous shales. Massive, gray, granular rocks, with prominent porphyroblasts of muscovite from 1 to 5 millimeters across, differ from the spangled muscovite schists primarily in having more quartz and secondarily in having less muscovite and in being more massive. The groundmass consists of biotite, muscovite, quartz, and some plagioclase. A mode is given in Table 6, column 10.

Quartzite.—The massive gray granular quartzites are composed of quartz, with some biotite and muscovite. Modes are given in Table 6, columns 11 and 11a. The grains are relatively small, minerals over 1 millimeter in diameter being comparatively rare. A chemical analysis is given in Table 10, column 9.

Garnet granulite.—Thin pink beds, seldom over 1 millimeter thick, are a rather distinctive feature of the paraschists, particularly near the Fitch formation. A microscopic study has not been made, but megascopic study indicates that the grain size is 0.1 millimeter, and that the essential minerals are garnet and quartz, with lesser amounts of biotite and muscovite. Hadley (1941) reports 5 per cent apatite from similar rocks in the Mt. Cube area.

MINERALOGY OF THE PARASCHISTS

For almost every specimen β, γ, and the optic angle of muscovite were obtained. The data are constant and indicate essentially pure $H_4K_2Al_6Si_6O_{24}$. An analysis of muscovite with nearly identical optical properties is given in Table 10, column 19. The γ index of biotite was obtained for nearly every specimen and the optic angle for some. Although the indices alone are not sufficient to determine the precise character, they indicate an intermediate biotite in which the phlogopite-eastonite series makes up 30 to 50 per cent and the annite-siderophyllite series make up to 70 to 50 per cent. Optical data on staurolite were obtained for only a limited number of specimens, but they are remarkably uniform. Moreover, they are very similar to the data for the staurolite from the Littleton-Moosilauke area, a chemical analysis of which is available (Table 10, column 17).

All of the chlorite is of retrograde origin, either replacing biotite or garnet. The optical data are rather uniform, and indicate an iron-rich end of the chlorite series, with 80 per cent of the iron molecule and only 20 per cent of the magnesian molecule (Winchell, 1933, p. 278). Oligoclase, ranging from 15 to 30 per cent anorthite, averages only 2 per cent of the formation.

The garnets are difficult to study, because they are rare, are always small grains, and generally have impurities. In no case is it possible to obtain garnet in sufficient quantities and proper purity for chemical analysis. Five rock specimens were studied. In material from a single hand specimen the index of refraction is constant, but the specific gravity varies, due to impurities, such as quartz, muscovite, and chlorite. Therefore, only the highest value for the specific gravity is significant. The figures range from 4.100 to 4.245, the average being 4,200; the average of the maximum value determined for each hand specimen is 4.230, and this is undoubtedly the more significant figure. The values for the indices of refraction range from 1.798 to 1.807, averaging 1.803.

From Winchell's (1933, p. 176) Figure 93 it is apparent that the garnets belong to the almandite-spessartite series, with perhaps limited amounts of andradite, grossularite, and pyrope in solid solution. Using his more detailed graphs, Figures 97, 98, and 99, the gravity readings suggest that the composition is near the almandite end of the series, whereas the indices are near the spessartite end. In the blowpipe studies the manganese reaction is weak to fair, indicating a limited content of manganese. The garnets of the Mt. Washington area are probably not very different from the analysed almandite from the Belknap Mountains (Table 10, column 16).

GNEISS

Locally the basal portion of the Littleton formation has been converted to gneiss. Three such areas are of sufficient size to show on the geological map (Pl. 1). One is 1½ miles north of Pinkham Notch, a second is 1 mile west of Mt. Clay, and a third is 1 mile east of Mt. Bowman. Moreover,

TABLE 6.—*Approximate modes of the Littleton formation*

Type	1	2	3	3a	4	5	6	7	8	9	10	11	11a	12
Number of thin sections	1	2	5	..	1	6	3	3	6	3	4	5	..	1
Quartz	48	37	23	29	46	45	57	39	52	55	66	87	72	50
Microcline	..	..	..	..	..	tr	..	..	..	..	..	..	..	..
Plagioclase	..	..	2	10	..	2	..	3	tr	tr	3	3	6	23
Biotite	20	17	20	24	15	15	24	9	21	4	14	8	13	1
Muscovite	20	27	40	21	20	25	12	15	20	23	16	2	4	..
Chlorite	..	7	4	..	..	3	..	13	1	13	1	tr	..	..
Sericite	..	9	4	..	8	6	..	16	1	..	..	tr	..	..
Andalusite	8	..	..	..	..	..	..	..	..	..	..	..	..	..
Sillimanite	..	2	3	11	7	2	..	2	2	..	tr	tr	4	..
Staurolite	..	..	1	2	1	tr	5	..	1	..	tr	..	..	..
Garnet	4	1	3	3	2	1	2	2	1	1	tr	tr	1	6
Magnetite	..	tr	tr	..	..	tr	..	1	1	tr	tr	tr	tr	2
Tourmaline	tr	tr	tr	..	1	tr	tr	tr	tr	1	tr	tr	..	tr
Pyrite	..	..	tr	..	..	1	..	tr	..	3	..	tr	..	3
Apatite	..	..	tr	tr	..	..	tr	tr	tr	tr	..	tr	tr	..
Zircon	..	..	..	..	..	..	..	..	tr	..	..	tr	..	..
Ilmenite	..	..	tr	..	..	..	..	..	tr	..	..	..	..	..
Hornblende	..	..	..	..	..	..	..	..	..	..	..	..	..	15
Carbon ?	..	..	..	..	..	..	1	..	..	..	..	..	..	..
Per cent of anorthite in Plagioclase:														
Average	..	..	20	5	..	22	..	18	..	..	15	12	8	90
Range	..	..	15–25	5	..	15–30	..	18	..	..	10–20	10–15	8	90
Grain size in millimeters	0.05–80.0	0.05–1.00	0.05–8.00	0.10–4.00	0.50–10.0	0.05–3.00	0.10–3.50	0.05–3.00	0.03–10.00	0.05–1.50	0.05–6.00	0.05–2.00	0.05–2.00	0.20–1.00

1. Coarse andalusite schist.
2. Coarse pseudo-andalusite schist.
3. Coarse rough pseudo-andalusite schist.
3a. Coarse rough pseudo-andalusite schist, calculated from chemical analysis, specimen W 223.
4. Intermediate pseudo-andalusite schist.
5. Shiny fine-grained pseudo-andalusite schist.
6. Staurolite schist.
7. Medium-grained massive schist.
8. Spangled muscovite schist.
9. Shiny fine-grained mica schist.
10. Mica-quartz schist.
11. Quartzite.
11a. Quartzite, calculated from chemical analysis, specimen W 222.
12. Lime-silicate concretion.

within the area shown as paraschist, there are numerous localities where the paraschists might be classed as incipient gneisses. Inasmuch as the gneisses do not differ from those in the Partridge formation, a separate description seems unnecessary.

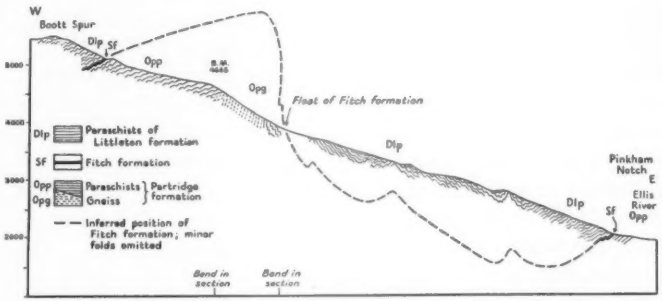


FIGURE 4.—*Structure section along ridge south of Cutler River*
Between Boott Spur and Pinkham Notch. Horizontal scale same as vertical.

THICKNESS

In many localities the Littleton formation has been so extensively folded that even approximate figures of the thickness are difficult to obtain. Elsewhere, the structure is comparatively simple, and reasonably accurate figures for part of the formation may be determined. On the ridge south of the Cutler River between altitudes of 2000 and 2700 feet, the lower part of the formation is 1200 feet thick (Fig. 4). Along the headwaters of Peabody River (Pl. 1) and the north fork of Cutler River the beds dip gently westward and 2000 feet of paraschists are exposed above the Fitch formation. On the valley slopes of the West Branch of the Peabody River and on Mt. Adams and Mt. Madison even higher strata in the Littleton are exposed, and structure sections indicate that the formation is about 4000 feet thick (Pl. 9, sections BB′, CC′, DD′).

OLIVERIAN MAGMA SERIES (DEVONIAN?)

GENERAL STATEMENT

The Oliverian magma series occupies the northwestern and northern parts of the area (Pl. 1), and extends far to the west, north, and northeast beyond the limits of the map. Inasmuch as these rocks are to be discussed fully in a forthcoming joint paper with R. W. Chapman and C. A. Chapman, detailed consideration here is not necessary. The following

types have been recognized: (1) biotite gneiss; (2) porphyritic biotite gneiss; (3) coarse granite; (4) coarse syenite; (5) hornblende-quartz monzonite; and (6) fine-grained gray monzonite.

AGE RELATIONS

The Oliverian magma series is younger than the Ammonoosuc volcanics. The relations of the biotite gneiss to the Ammonoosuc volcanics are best displayed 2½ miles west-southwest of Randolph on Cold Brook, not shown on Plate 1. Between altitudes of 1400 and 1480 feet biotite gneiss of the Oliverian series contains inclusions of amphibolite from the Ammonoosuc volcanics. At 1490 feet, the main body of the Ammonoosuc volcanics begins, but there is at least one dike of biotite gneiss.

Two and one-half miles west of the summit of Mt. Deception, near a narrow belt of Ammonoosuc volcanics, reworked inclusions of the volcanics are common in the porphyritic biotite gneiss; moreover, dikes and sills of the porphyritic biotite gneiss intrude the volcanics.

Although in the Mt. Washington area it is impossible to determine the age of the Oliverian magma series relative to the formations younger than the Ordovician (?), in the Mascoma quadrangle C. A. Chapman (1939, p. 168) has shown that the Oliverian series intrudes the Silurian Clough formation. On structural evidence Billings (1937, p. 502) as well as C. A. Chapman and J. B. Hadley (1939) consider the Oliverian to be younger than the Devonian Littleton formation.

BICKFORD GRANITE (DEVONIAN?)

GENERAL STATEMENT

Many square miles of the Mt. Washington area are underlain by the Bickford granite. Four principal bodies have been distinguished. The Bretton Woods body is the largest and lies west of Mt. Monroe, extending from the lower slopes of the Presidential Range west to Bretton Woods. A large appendage extends northeasterly beyond Jefferson Notch. The second area is 1¼ miles southeast of Mt. Bowman. A third, irregular area lies south of Randolph, and a fourth, east of Mt. Madison, centers about Dolly Copp Camp.

Accurately portrayal of the shapes of these bodies of Bickford granite is difficult, partly because of the lack of exposures in some places. This is notably true of the Randolph body, where large areas are devoid of outcrops. Another difficulty arises in large areas where the older schists are cut by numerous dikes and sills of Bickford granite, south of Mt. Deception and Mt. Dartmouth. Wherever exposures are poor it is difficult to decide whether an outcrop of granite belongs to a large body or is

merely a sill or dike in a much larger area of gneiss or schist. Thus the contact around Mt. Dartmouth and Mt. Mitten is poorly defined. One and a half miles southwest of Randolph, outcrops of Bickford granite alternate with the Ammonoosuc volcanics. On the geological map these outcrops have been arbitrarily joined to the main part of the Randolph body.

Good and accessible exposures may be seen at the following localities: (1) Ammonoosuc River, upstream from Bretton Woods; (2) Ammonoosuc River, between 2080 and 2260 feet; (3) the trail on the south slope of Little Mt. Deception.

LITHOLOGY

Megascopic description.—The Bickford is a white to gray binary granite (Pl. 6, fig. 4). The texture is equigranular, locally porphyritic, and platy textures are rare. The chief minerals are feldspar, quartz, muscovite, and biotite. The white subhedral to anhedral feldspar grains show Carlsbad twins and, less commonly, albite twins. Quartz is milky white. Biotite and muscovite constitute 2 to 7 per cent each.

The only significant difference between specimens of the Bickford granite is the grain size. In the coarsest specimen collected the feldspar crystals range from 3 to 15 millimeters in length, the quartz and micas 1 to 3 millimeters. In the finest grained specimens the individual grains range from 0.3 to 1.0 millimeter. All transitions between these extremes have been observed. The coarser varieties are characteristic of the larger bodies, the finer are found in the small bodies.

Microscopic description.—The essential minerals of the Bickford granite are quartz, potash feldspar, oligoclase, biotite, and muscovite, with such accessories as apatite and garnet. The quartz in many cases has undulatory extinction. Carlsbad twinning is particularly noticeable in the larger crystals of potash feldspar. Most grains have the cross-twinning characteristic of microcline. The oligoclase has both Carlsbad and albite twinning, and some crystals are weakly zoned with more calcic cores.

The average of 10 modes given in Table 7, column 1, is quartz monzonite, but lies so near the borderline with granite, that it seems advisable to call the whole unit granite. Some specimens are quartz monzonites; others are granites.

As Table 7 shows, the anorthite content of the plagioclase is relatively uniform, ranging in different specimens from 12 to 24 per cent, and averaging 17.5 per cent. The muscovite is relatively uniform, with γ averaging 1.601 and ranging from 1.599 to 1.604; 2V averages 36 degrees, with a range from 34 to 37½ degrees. For biotite γ averages 1.651, with a range of 1.645 to 1.658; 2V is generally small but in one case is 21 degrees.

Under the microscope the Bickford granite shows a poorly defined hypidiomorphic granular texture, with strong granulitic tendencies. Many grains of plagioclase, potash feldspar, and mica are subhedral, but other grains of these minerals and all the quartz are distinctly anhedral. A little myrmekitic intergrowth of quartz and plagioclase was noted in several specimens.

AGE RELATIONS

The Bickford granite is younger than the metamorphic rocks and the Oliverian magma series but is older than the White Mountain magma series. Being undeformed, it must be younger than the main orogenic

TABLE 7.—*Approximate modes—Bickford granite*

	1	2	3	4
Number of thin sections	10	1	1	1
Quartz	30	42	35	40
Microcline*	38	....	39	....
Plagioclase	24	35	20	....
Muscovite	4	17	2	5
Biotite	4	2	4	....
Apatite	tr	tr	tr	....
Garnet	tr	....	....	....
Chlorite	....	4	....	....
Sericite	....	....	....	55
Leucoxene	....	....	....	tr
Per cent of anorthite in plagioclase:				
Average	17.5	16	16	....
Range	12–24	....	....	....
Grain size in millimeters	0.1–15.0	0.5–3.0	0.1–1.0	

* Includes orthoclase.

1. Bickford granite.
2. Binary granite, A. M. C. bridge over Cutler River.
3. Binary granite, north slope of Pine Peak, 2400 feet.
4. Granitic rock, adjoining shear zone and showing effect of shear, from top of 2980 knob, approximately 1 mile SW. of Pine Peak.

period. It is the youngest member of the New Hampshire magma series, but has been silicified and sericitized near the Pine Mountain fault.

Dikes and sills of Bickford granite cut the Ammonoosuc volcanics on several of the brooks 1 to 2 miles southwest of Randolph. On Plate 1 a special symbol is used for those areas where dikes and sills of the Bickford granite cut the Partridge and Littleton formations. The Oliverian magma series, being more deformed and granulated, is believed to be older than the Bickford granite. Moreover, for reasons discussed elsewhere (Billings, 1937, p. 535-536) the Oliverian series is considered to be older than the folding.

No data are available in this area bearing on the age of the Bickford granite relative to the syenite of Cherry Mountain and the Conway granite. However, blocks of binary granite similar to the Bickford have been found in the Moat volcanics (Billings, 1928, p. 99-100), which are older than the syenites and the Conway granite of the White Mountain

magma series (Billings, 1928, p. 88). Hence the Bickford granite is considered older than the White Mountain magma series.

WHITE MOUNTAIN MAGMA SERIES (MISSISSIPPIAN?)

The Mississippian (?) White Mountain magma series is represented by the syenite at Cherry Mountain (Mt. Martha), the Conway granite in the southwest corner of the map (Pl. 1), and small volcanic vents composed of tuff, breccia, and diabase. R. W. Chapman (1937) has given a complete description of the syenite so that further discussion is unnecessary. The Conway granite is like that in the adjacent Franconia quadrangle (Williams and Billings, 1938). The volcanic vents will be considered in a forthcoming paper by K. F. Billings describing the dike rocks.

CORRELATIONS AND AGE OF FORMATIONS

GENERAL STATEMENT

In the preceding pages the lithologic units have been described under names based on correlation with the Littleton-Moosilauke area (Billings, 1937) and elsewhere. No proof of the correlations adopted or of the ages assigned has been given.

In the Mt. Washington area certain standard methods of correlation can not be used. Paleontological methods fail, as fossils have not been found. The writer believed that possibly one or more stratigraphic units could be followed northeasterly from the Littleton-Moosilauke area to the Mt. Washington area (Fig. 1). The western half of the area covered by Plate 1 was mapped chiefly for that purpose. As the map shows, the project failed because the southwest corner of the Mt. Washington area is underlain by plutonic rocks. The Oliverian magma series, however, can be followed from the Littleton-Moosilauke area to the Mt. Washington area.

Correlation between these areas must be based on lithologic similarity and a similar sequence of rock types, including the thickness of the units involved. The correlation is made with the Littleton-Moosilauke area because that is the only place in New Hampshire where fossils—middle Silurian and lower Devonian—have been found. Moreover, the stratigraphic sequence was first established there (Billings, 1937). The rocks in that locality show a progressive increase in metamorphism toward the southeast and were classified as belonging to the low-grade, middle-grade, and high-grade metamorphic zones.

LITHOLOGIC SIMILARITY

Ammonoosuc volcanics.—The thick sequence of metamorphosed volcanics on the northern slopes of the Presidential Range was correlated

with the Ammonoosuc volcanics early in the field work. Not only are the rocks similar to those found in the Littleton-Moosilauke area, but they are very thick. Although volcanics similar to those in the Ammonoosuc formation are found in the Littleton formation in the Littleton and Moosilauke quadrangles, they are comparatively thin. Moreover, the

TABLE 8.—*Stratigraphic column of the Littleton-Moosilauke area*

AGE	FORMATION	INFERRED ORIGINAL LITHOLOGY	THICKNESS (feet)
Lower Devonian	Littleton formation	Sandstone and shale, with some thin volcanic members	5000
Middle Silurian	Fitch formation	Limestone, arenaceous limestone, dolomitic shale, arenaceous dolomite, dolomitic sandstone, calcareous shale, sandstone, shale	400–700
Lower or middle Silurian	Clough formation	Quartz conglomerate and sandstone	0–150
Upper Ordovician (?)	Partridge formation	Shale and sandstone	0–2000
	Ammonoosuc volcanics	Tuffs, breccias, and conglomerates, with only a few flows, of the composition of soda-rhyolite (quartz latite), andesite, basalt	2000±
	Albee formation	Sandstone and shale	4000±

metamorphosed volcanics in the Mt. Washington area occupy the same structural position as the Ammonoosuc volcanics in the Franconia and Moosilauke quadrangles. In both localities they form the roof of an intrusive dome of the Oliverian magma series. Although it is impossible to trace the Ammonoosuc volcanics from one area to another, the outcrops are on the same strike. A small area of Ammonoosuc volcanics 3 miles north of Mt. Oscar (Pl. 1) lies between the two localities.

Fitch formation.—The Fitch formation is unique in west-central New Hampshire because of its original content of calcareous and dolomitic beds. In the higher metamorphic zones of the type locality these rocks contain such distinctive minerals as diopside, actinolite, calcic plagioclase, magnesian biotite, and sphene. Similar rocks are unusual and striking in the monotonous succession of schists of the Mt. Washington area. When they were first observed their similarity to the Fitch formation of the Littleton-Moosilauke area was at once recognized. One difference, however, should be noted: marbles are characteristic, although not

abundant in the Fitch formation of the Littleton-Moosilauke area, but they are absent from the Mt. Washington area.

Littleton formation.—The paraschists of the Presidential Range are lithologically similar in most respects to the Littleton formation around Mt. Moosilauke. In the Mt. Washington area, however, large andalusite

TABLE 9.—*Stratigraphic sequence in the Mt. Washington area.*

UNIT	ORIGINAL LITHOLOGY	THICKNESS (feet)
d.	Sandstone and shale	4000
c.	Dolomitic shale, dolomitic sandstone, arenaceous dolomite, sandstone, shale, calcareous shale	0–200
b.	Shale and sandstone	1400±
a.	Tuffs—and perhaps breccias and conglomerates—with the composition of quartz latites, rhyolites, andesite, and basalt	5000±

and pseudo-andalusite crystals are prominent, but this is due to a difference in the metamorphic history. The original sediments were shales and sandstones in both areas.

SIMILAR SEQUENCE

Correlations based on lithologic similarity alone are not necessarily convincing. However, the sequence in the Mt. Washington area is similar to that in the Littleton-Moosilauke area, the sequence for which is summarized in Table 8. The lithology is given in terms of the original sediments, and, although inferred, is essentially the same as that found in the low-grade zone. The thicknesses are also given.

Table 9 gives the sequence in the Mt. Washington area, the lithology likewise in terms of the original sediments. The columns in Tables 8 and 9 match satisfactorily. At the top of both columns is a thick series of shale and sandstone, and "d" of Table 9 corresponds to the Littleton formation of Table 8. Next lower in the Mt. Washington area is a comparatively thin series of metamorphosed dolomitic and calcareous beds, which correspond to the Fitch formation.

The Clough formation is absent from the Mt. Washington area, but this situation is expectable. This formation is not present in the northeast portion of the Littleton-Moosilauke area, "for the formation was deposited as a thin sheet overlapping from the southwest" (Billings, 1937, p. 48). The thinness of the Fitch formation in the Mt. Washington area com-

pared with the Littleton-Moosilauke area, is presumably due to the same overlapping of the Silurian seas from the southwest.

Unit "b" of the Mount Washington area agrees in position, lithology, and thickness with the Partridge formation. Finally, unit "a" of Table 9 agrees in position and lithology with the Ammonoosuc volcanics, although it is considerably thicker around Mount Washington.

PLUTONIC ROCKS

Oliverian magma series.—The correlation of this series is based on several facts: (1) some units are mineralogically and texturally similar to Oliverian rocks elsewhere; (2) the attitude of the foliation indicates a great dome, as in the type locality of the Oliverian series; and (3) this dome is located in the center of the Bronson Hill anticline, a feature characteristic of the Oliverian series elsewhere.

Bickford granite.—The binary granite south and east of the Dartmouth Range is similar to the Bickford granite of the type locality in the Franconia quadrangle (Williams and Billings, 1938). These rocks also resemble the Concord granite at Concord, New Hampshire, and Hitchcock (1877, p. 161) applied this name in the Mt. Washington area. However, this distant correlation may be incorrect, so that it seems advisable to use a name from a nearby locality. The term Randolph granite (Billings, 1928, p. 81) was also used for these rocks but should be abandoned because of prior use elsewhere (Wilmarth, 1938, p. 1770-1771).

AGE

The geologic age of the lithologic units has been discussed in previous papers, so that a thorough consideration is unnecessary. Fossils from the vicinity of Littleton, New Hampshire, place the Fitch formation in the middle Silurian and the Littleton formation in the lower Devonian (Billings and Cleaves, 1934). The Ammonoosuc volcanics and the Partridge formation are probably upper Ordovician, as they lie unconformably beneath the Silurian and are apparently above the middle Ordovician of Vermont (Billings, 1937, p. 475). The Oliverian magma series and the Bickford granite are younger than the lower Devonian and are probably late Devonian. The White Mountains magma series is tentatively considered Mississippian.

COMPARATIVE MINERALOGY

GENERAL STATEMENT

Special attention was given in the present study to the chemical composition of the minerals in the metamorphic and igneous rocks. This was deduced from optical data. No new analyses of minerals were made, for

TABLE 10.—*Chemical analyses of rocks a*

	1	2	3	4	5	6	7	8	9	10	1
SiO_2.....	58.92	58.14	66.36	61.40	56.23	66.68	57.43	86.91	82.56	62.87	50.
TiO_2.....	0.93	0.65	1.03	0.92	1.11	0.87	0.94	0.50	0.64	0.92	1.
Al_2O_3.....	18.55	21.00	15.75	19.04	23.15	18.17	21.68	6.34	8.22	17.43	16.
Fe_2O_3.....	0.94	0.33	1.16	0.63	1.17	0.90	0.08	0.59	0.50	0.55	1.
FeO.....	6.63	6.32	5.37	6.90	6.94	4.98	6.82	2.03	3.22	6.67	8.
MnO.....	0.08	0.06	0.08	0.07	0.12	tr	0.28	0.02	0.04	0.23	0.
MgO.....	3.24	3.41	2.50	2.10	2.21	1.42	2.02	0.58	1.02	2.71	6.
CaO.....	0.48	0.32	0.52	0.90	0.26	0.76	0.72	0.08	0.26	0.59	9.
Na_2O.....	1.49	1.10	1.68	1.30	1.14	0.74	0.53	0.14	0.72	1.71	2.
K_2O.....	3.74	3.85	3.23	3.74	4.53	2.89	6.32	1.32	1.69	4.08	0.
H_2O+.....	3.90	4.47	1.95	2.89	2.75	1.75	2.80	1.23	0.91	2.10	0.
H_2O-.....	0.11	n.d.	0.06	0.06	0.09	0.40	0.20	0.08	0.05	0.08	0.
CO_2.....	0.25	0.00	0.01	0.01	0.01	0.00	0.00	0.00	0.01	0.02	0.
ZrO_2.....	n.d.	0.04	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.
P_2O_5.....	0.14	0.00	0.17	0.08	0.18	tr	tr	0.04	0.07	0.11	0.
S.....	0.17	0.09	0.03	0.08	0.02	0.05	tr	0.05	0.02	0.04	0.
C.....	n.d.	n.d.	n.d.	n.d.	n.d.	0.29	0.00	n.d.	n.d.	n.d.	n.
BaO.....	n.d.	0.00	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.
O−S Corr..	−0.06	.....	−0.01	−0.03	−0.01	.....	.....	−0.02	−0.01	−0.02	−0.
Total.....	99.51	99.78	99.89	100.09	99.90	99.90	99.82	99.94	99.92	100.09	99.
S. G.....	2.75	2.777	2.83	2.75	2.91	.....	.....	2.69	2.72	2.83	2.

* Not included in Daly's average analyses.

1. (W 219) Slate, Littleton formation, low-grade zone, quarry at Slate Ledge, approximately 2¾ miles west of Littleton, N. H. "Several of your samples, contain appreciable amounts of graphitic carbon, especially the slate, W 219. This probably accounts for the rather low summation of that analysis." (Letter from R. B. Ellestad, May 8, 1940). G. Kahn, analyst.

2. (L 489) Slate, Littleton formation, low-grade zone, quarry at Slate Ledge, approximately 2¾ miles west of Littleton, N. H. (Billings, 1937, p. 556). F. A. Gonyer, analyst.

3. (L 494) Biotite-garnet-staurolite schist, Littleton formation, middle-grade zone, 0.5 mile south of Northey Hill, Moosilauke quadrangle. "Some difficulty was encountered with sample L 494, the staurolite schist, in connection with the ferrous iron determination. This mineral is extremely resistant to the HF-H_2SO_4 attack used in the usual method for FeO determinations in silicates. Consequently, in the figures reported, the FeO in the staurolite would be counted as ferric oxide. In order to get an idea of the error involved, I determined the total iron content of the residue left after the usual FeO determination, and found 0.62% Fe_2O_3. If this were all present as FeO originally, the reported figure for Fe_2O_3 would be high by 0.62%, while the reported figure for FeO would be low by 0.62×0.9, or 0.56%." (Letter from R. B. Ellestad, May 17, 1940). G. Kahn, analyst.

4. (W 229) Shiny fine-grained pseudo-andalusite schist, Partridge formation, high-grade zone, 2.5 miles S. 75° E. from Mt. Washington (Cutler River, 2135 feet). G. Kahn, analyst.

5. (W 223) Coarse rough pseudo-andalusite schist, Littleton formation, high-grade zone, 2.1 miles N. 32° E. from Mt. Washington (Chandler Ridge, 4060 feet). G. Kahn, analyst.

6. Sillimanite schist, Littleton formation, high-grade zone, 0.2 mile NW. of summit of South Peak of Loon Mountain, Franconia quadrangle, N. H. (Billings, 1938a, p. 232). W. H. Herdsman, analyst.

7. Muscovit
Mountain, Fr
8. (W 230)
Littleton qua
9. (W 222)
Washington
10. (W 215)
Washington
11. (W 56)
Bowman, Mt.
12. (L 416)
zone, 1 mile
556). F. A.
13. Average
14. (W 26)
Mt. Washing
15. Average
16. Garnet
tains area.
17. Staurol
of Garnet H
18. Commo
Moosilauke
19. Muscov

9	10	11	12	13	14	15	16	17	18	19
82.56	62.87	50.91	51.64	49.06	75.93	72.80	36.21	27.81	44.18	46.31
0.64	0.92	1.68	1.33	1.36	0.18	0.33	n.d.	n.d.	1.23	0.79
8.22	17.43	16.00	17.62	15.20	12.61	13.49	21.32	54.09	10.34	38.52
0.50	0.55	1.17	1.14	5.38	0.21	1.45	3.80	2.76	3.02	tr
3.22	6.67	8.81	7.80	6.37	1.13	0.88	31.20	12.48	13.90	n.d.
0.04	0.23	0.21	0.12	0.31	0.03	0.08	5.12	n.d.	tr	n.d.
1.02	2.71	6.85	7.74	6.17	0.58	0.38	1.41	1.92	11.39	0.18
0.26	0.59	9.99	6.44	8.95	0.38	1.20	0.36	n.d.	12.31	0.46
0.72	1.71	2.27	4.52	3.11	2.76	3.38	n.d.	n.d.	1.11	1.56
1.69	4.08	0.69	0.25	1.52	5.87	4.46	n.d.	n.d.	0.82	8.03
0.91	2.10	0.97	1.29	1.62	0.21	1.47	n.d.	1.70	1.57	4.46
0.05	0.08	0.05	n.d.		0.01		n.d.	n.d.	n.d.	
0.01	0.02	0.07	0.00	*	0.00	*	n.d.	n.d.	0.00	n.d.
n.d.	n.d.	n.d.	0.00	*	n.d.	*	n.d.	n.d.	n.d.	n.d.
0.07	0.11	0.20	0.09	0.45	0.02	0.08	n.d.	n.d.	0.00	n.d.
0.02	0.04	0.03	tr	*	0.02	*	n.d.	n.d.	0.00	n.d.
n.d.	n.d.	n.d.	n.d.	*	n.d.	*	n.d.	n.d.	n.d.	n.d.
n.d.	n.d.	n.d.	0.00	*	n.d.	*	n.d.	n.d.	n.d.	n.d.
−0.01	−0.02	−0.01	.....	.....	−0.01	.....	.....	.....	.....	.....
99.92	100.09	99.89	99.98	100.00	99.93	100.00	99.42	100.76	99.86	100.31
2.72	2.83	2.92	2.901	.....	2.61	.....	4.23	3.775	.....	.....

7. Muscovitized schist, Littleton formation, high-grade zone, summit of South Peak of Loon Mountain, Franconia quadrangle, N. H. (Billings, 1938a, p. 292). W. H. Herdsman, analyst.

8. (W 220) Sandstone, Littleton formation, low-grade zone, 0.3 mile SE. of quarry at Slate Ledge, Littleton quadrangle, N. H. G. Kahn, analyst.

9. (W 222) Quartsite, Littleton formation, high-grade zone, 2.1 miles N. 32° E. from Mt. Washington (Chandler Ridge, 4000 feet). G. Kahn, analyst.

10. (W 215) Gneiss, Partridge formation, high-grade zone, 1.6 miles N. 5° E. from Glen House, Mt. Washington area. G. Kahn, analyst.

11. (W 56) Amphibolite, Ammonoosuc volcanics, high-grade zone, 0.2 mile S. 65° E. from Bowman, Mt. Washington area. G. Kahn, analyst.

12. (L 416) Fine-grained amphibolite, volcanic member of the Littleton formation, middle-grade zone, 1 mile above the mouth of the Gale River, Littleton quadrangle, N. H. (Billings, 1937, p. 556). F. A. Gonyer, analyst.

13. Average basalt, analysis No. 58, from Daly (1933, p. 17).

14. (W 26) Biotite gneiss, Ammonoosuc volcanics, high-grade zone, 0.6 mile south of Bowman, Mt. Washington area. G. Kahn, analyst.

15. Average rhyolite, analysis No. 5, from Daly, 1933, p. 9.

16. Garnet from mica schist, west slope of Piper Mountain, altitude 1300 feet, Belknap Mountains area. (Modell, 1936, p. 1006). F. A. Gonyer, analyst.

17. Staurolite from Littleton formation, middle-grade zone, probably from 0.8 mile due west of Garnet Hill, Moosilauke quadrangle. S. L. Penfield and J. H. Pratt (1894).

18. Common hornblende from amphibolite sill, 0.6 mile S. 75° E. from summit of Garnet Hill, Moosilauke quadrangle. (Billings, 1937, p. 556).

19. Muscovite from high-grade schist, Clove quadrangle, New York (Barth, 1936, p. 780).

in many cases the relation between optics and chemical composition is known. In other cases, such as the biotite group, where the relation is not so definite, new analyses did not seem advisable. Even under the best circumstances a certain amount of contamination occurs, and the presentation of analyses of such material would be misleading.

Optical data, including indices, were obtained on all the essential minerals in all the rocks for which thin sections were available. For comparison such data were plotted on graphs. The indices of refraction are the most significant data, and γ is the abscissa in Plate 8. Each lithologic unit is shown by a distinctive line. The ordinate could be based on the total number of determinations, but such a method would not be useful for comparative purposes. The ordinate represents, therefore, the percentage of the total number of readings falling within a given range of indices. The range chosen was 0.005. For example, in the plagioclase diagram, the curve for the Fitch formation reaches 55 per cent at the value of 1.6475. This means that of all the determinations made, 55 per cent fell between 1.645 and 1.650.

In some cases, notably plagioclase and the actinolite-tremolite series, where only two end members are involved, it is possible to relate the curves directly to chemical composition, as has been done in Plate 8.

The graphs serve two purposes. They show the extent to which one mineral varies in composition within a single unit and also how much one mineral varies from unit to unit.

COMPOSITION OF MINERALS

Plagioclase.—The composition of the plagioclase differs between formations. In the paraschists of the Littleton and Partridge formations 60 per cent of the plagioclase is close to An_{10}, and none is more calcic than An_{25}. In the gneisses of these same formations the plagioclase is more calcic; the maximum on the curve is at An_{22} and none is more calcic than An_{45}. The plagioclase in the Fitch formation is very calcic in many specimens. The maximum is near An_{90}, but there is a distinct submaximum at An_{30}.

In the Ammonoosuc volcanics, which in this area consist almost exclusively of biotite gneisses and related types derived from rhyolite and quartz latites, the crest of the curve is at An_{45}. None is more calcic than An_{55}, but some are very sodic.

In the Bickford granite the maximum covers the range from An_{10} to An_{22}, and varieties as calcic as An_{35} are unknown.

In order of increasing anorthite content of the maximum the lithologic units are, as shown by Plate 8: paraschists of the Littleton and Partridge formations (An_{10}), Bickford granite (An_{10} to An_{22}), gneisses of the Littleton and Partridge formations (An_{30}), Ammonoosuc volcanics (An_{45}), and the Fitch formation (An_{90}).

Muscovite.—The muscovite in the Mt. Washington area has constant optical properties. For the Bickford granite and the paraschists and gneisses of the Littleton and Partridge formations, γ in more than 75 per cent of the specimens lies between 1.600 and 1.605. The optic angle was measured for every hand specimen and the

average value for 2V is 37½ degrees. Barth (1936, p. 780) gives an analysis (Table 10, column 19 of the present paper) of muscovite with similar optical properties.

In the Ammonoosuc volcanics the maximum is a little higher, between 1.605 and 1.610; 2V averages 37½ degrees. The composition would differ very slightly from the other formations.

Biotite.—The maximum for γ of biotite from the Fitch formation is much lower than for the other units. In nearly 50 per cent of the specimens γ lies between 1.610 and 1.620. This indicates that the biotite is nearer the magnesian than the iron end of the series, and that phlogopite-eastonite molecules constitute approximately 65 per cent of the mineral.

For biotite in the Bickford granite, as well as the paraschists and gneisses of the Littleton and Partridge formations, γ reaches a maximum between 1.640 and 1.650. These data indicate that the biotite is nearer the iron end of the series, suggesting approximately 60 per cent of annite-siderophyllite.

Amphibole.—Amphibole occurs abundantly only in the Fitch formation, and to a lesser extent in the Ammonoosuc volcanics. In the Fitch formation over 50 per cent of the amphibole has γ close to 1.647, and the remainder does not differ greatly. Complete optical data show that the amphibole is actinolite with about 80 per cent of the tremolite molecule, and about 20 per cent of the $CaFe_3Si_4O_{12}$ molecule (Winchell, 1933, p. 246). The γ index of the amphibole in the Ammonoosuc volcanics is on the average distinctly higher than in the Fitch formation and covers a range from 1.663 to 1.701. Complete data indicate that it is common hornblende.

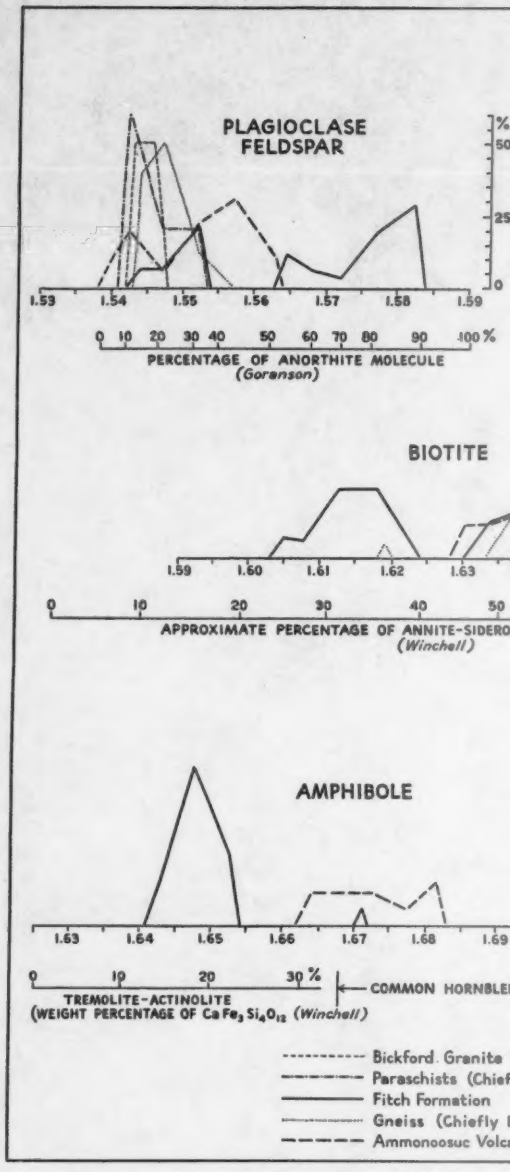
Pyroxene.—Pyroxene occurs only in the Fitch formation and the Ammonoosuc volcanics. In the Fitch formation γ is very uniform, more than 75 per cent of the specimens lying between 1.705 and 1.710. Complete data indicate diopside-hedenbergite with about 20 per cent hedenbergite and 80 per cent diopside (Winchell, 1933, p. 226).

Pyroxene is not abundant in the Ammonoosuc volcanics. It is distinctly richer in iron than in the Fitch formation, and the average content of hedenbergite is 55 per cent. The range is from 50 to 70 per cent.

The data concerning diopside and actinolite in the Fitch formation are significant for a number of reasons. They show that within the Mt. Washington area these two minerals have a rather limited range in composition, and indicate that the amount of the ferrous molecule in solid solution with the magnesia molecule is essentially the same for both minerals. Moreover, the data are the same as for the middle-grade lime-silicate rocks of the Fitch formation in the Littleton-Moosilauke area (Billings, 1937, p. 486), where the diopside carries 20 per cent of hedenbergite in solid solution and the actinolite 20 per cent $CaFe_3Si_4O_{12}$.

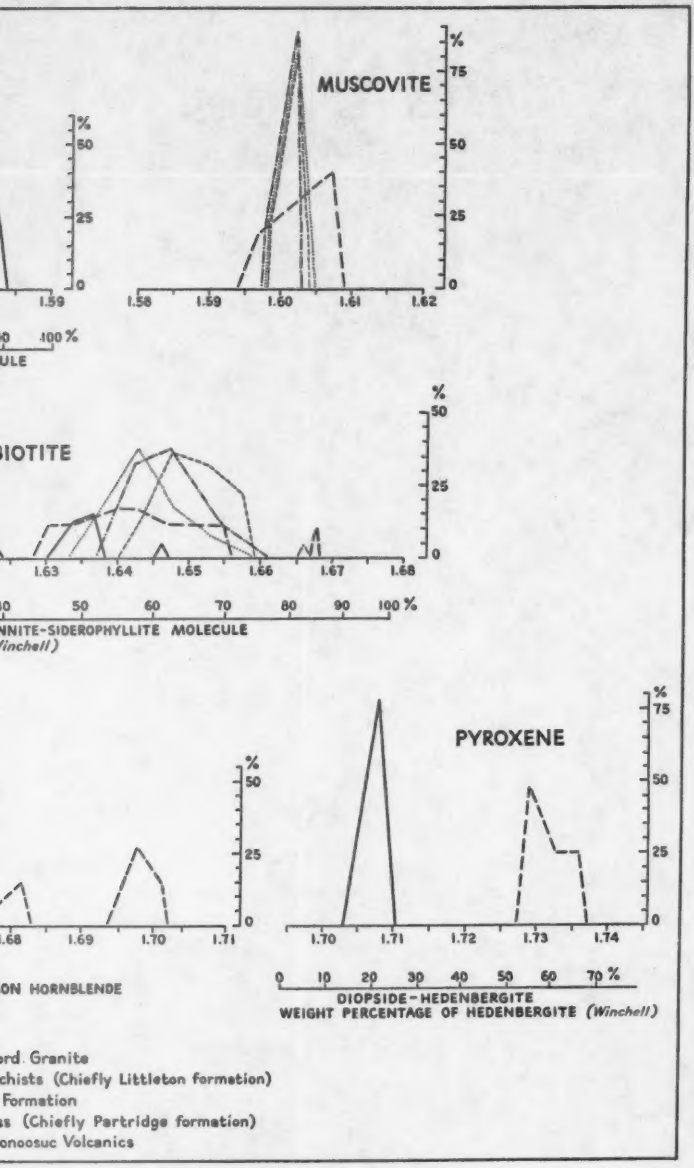
Garnet.—Although garnet is found in several of the lithologic units, it is nowhere abundant, and significant data are difficult to obtain. In the paraschists of the Littleton formation the garnet is almandite with considerable spessartite in solid solution.

Staurolite, sillimanite, and andalusite.—Andalusite and staurolite are largely confined to the paraschists of the Littleton formation, although some andalusite is also found in the Partridge formation. Staurolite has uniform optics and hence a constant composition (Table 10, column 17). No special optical study was made of the andalusite. Sillimanite is found in the paraschists and gneisses of the Partridge and Littleton formation, as well as a small body of gneiss in the Ammonoosuc volcanics. No special optical study was made.



COMPARAT

In all cases the abscissa is the γ index, the ordinate th percentage of the determinations falling within each 0.00 block. Beneath the abscissa in most cases there is also second scale to show the approximate composition of th



ord Granite
chists (Chiefly Littleton formation)
Formation
s (Chiefly Partridge formation)
onoosuc Volcanics

OMPARATIVE MINERALOGY

ordinate the n each 0.005 ere is also a sition of the mineral. The graphs show how each mineral varies within a given formation and also how each mineral varies from one formation to another.

Potash feldspar.—Potash feldspar is rare in the paraschists and gneisses of the Littleton and Partridge formations, having been observed in only one specimen. This is striking because in certain areas such as Dutchess county, New York, potash feldspar becomes progressively more abundant with an increase in metamorphism (Barth, 1936, p. 825-832). Apparently in the Mt. Washington area not enough potash was introduced to change the muscovite to potash feldspar.

The relative abundance of microcline in the Fitch formation contrasts with the lack of potash feldspar in the aluminous sediments. Some of the microcline may be recrystallized detrital feldspar. Much of it, however, may be due to the reaction of sericite and dolomite, forming microcline, actinolite, and anorthite.

Potash feldspar, chiefly microcline, is characteristic of the Ammonoosuc volcanics. As in the Littleton-Moosilauke area, it probably carries considerable albite in solid solution, but no optical study was made. Potash feldspar, chiefly microcline, is also common in the Bickford granite.

SUMMARY

Many of the minerals have relatively uniform optical properties and chemical compositions for a given stratigraphic unit but show rather systematic differences between units. Plagioclase, amphibole, and pyroxene are good examples. Biotite is a fair example, but muscovite is relatively constant throughout all the rocks.

STRUCTURE

GENERAL STATEMENT

The structure of the area is characterized by folds, faults, and intrusions.

In the Presidential Range major and minor asymmetrical, plunging en echelon folds dominate the structure. Trending at right angles to the axes of the folds are culminations and depressions; the folds plunge away from the culminations and toward the depressions. Faults are subordinate, and intrusions, except for small dikes and sills, are confined to the lower northern and western slopes of the range. In the Dartmouth Range the metamorphosed sediments are obviously folded, but details of the structure have not been solved.

The northeasterly trending Pine Mountain fault crosses the northwestern slopes of the Dartmouth Range and probably continues to a point north of Randolph. It is apparently a gravity (normal) fault with the downthrow on the southeast. Some of the many minor faults are shown on the geological map (Pl. 1).

Large igneous intrusions occupy the northwestern, western, and southwestern parts of the Mt. Washington area. The rocks of the Oliverian magma series in the Mt. Washington area are generally foliated, and although several different lithologic units are represented, the mass as a whole shows a distinct anticlinal structure. The New Hampshire magma series is represented by the Bickford granite, a massive binary granite,

rarely foliated, and in part, at least, cross-cutting. The White Mountain magma series is represented by parts of two stocks, the Cherry Mountain stock in the west, and a body of Conway granite in the southwest.

JEFFERSON DOME

The foliation of the Oliverian magma series shows a major anticline (Pl. 10; Pl. 9, section HH′), called the Jefferson dome, from the township of that name. The crest of this anticline lies 1.8 miles N. 30° E. from the summit of Hardwood Ridge (Pl. 10). Northwest of the anticlinal axis the foliation dips 10° to 30° NW., whereas south of the axis the dip is 16° to 30° SE. The intrusive rocks on the southeast limb have a breadth of 3 miles and dip southeast at angles of 30 to 90 degrees. The variations in dip are probably due to rolls in a southeasterly-dipping limb; there is no evidence of closed, isoclinal folding. From north of Hardwood Ridge, the axis must extend west-southwest toward Mt. Martha. The Oliverian magma series north of the syenite stock on Mount Martha is obviously part of the northern limb of the anticline. The occasional steep northerly dips southeast of the syenite are presumably due to local overturning of a steeply dipping limb. The symbols east of the syenite stock indicate deformation of the foliation. These exposures are probably near the crest of the anticline. The deformation may be older than the intrusion of the syenite stock or contemporaneous with it.

PRESIDENTIAL RANGE

Minor folds.—The Presidential Range is replete with minor folds. Although in many outcrops the attitude of the bedding is relatively constant, in others it has been contorted into small folds, the amplitude and wave length of which range from a few inches to scores of feet. Under favorable conditions such folds may be photographed (Pl. 7), otherwise small sketch maps can be prepared (Fig. 9).

In recording such folds, the attitude of two features, the axial plane and the axis must be measured. The strike and dip of the axial plane are recorded, as well as the strike of the horizontal projection of the axis and the plunge of the axis. In most papers only the attitude of the axis is recorded, but this information alone is insufficient to define the fold. Suppose the horizontal projection of the axis strikes north and the plunge is zero. A fold with a vertical axial plane striking north could satisfy this condition; or asymmetrical and overturned folds with the axial plane dipping at any angle to either the east or west could satisfy the conditions. It is apparent, therefore, that the attitude of both the axial plane and the axis should be recorded. The objection may be made that the axial plane is an imaginary surface and its attitude can not be properly measured.

Even though precise measurements may be difficult, the structural picture is incomplete without such information.

The attitude of the minor folds is given on Plate 10. In many places where more data were obtained than could be shown on the map, the symbol is an average of several folds. In the Presidential Range the axial planes usually dip steeply to the northwest. However, in an area that extends from 1 mile southwest of Pinkham Notch as far north as the Auto Road (Pl. 10) the axial planes dip steeply to the east. Moreover, on the northwest flanks of the Presidential Range the axial planes commonly dip to the southeast.

The direction of plunge of the minor folds is more variable. Nevertheless, over large areas the plunges are systematic both in direction and value. Between Pinkham Notch and Mt. Monroe gentle northerly plunges prevail. Around the summit of Mt. Washington the plunges are southerly, but on the north slope of Mt. Clay they are northerly. Near the Halfway House the minor folds plunge toward the north at high angles, but to the south the plunge is nearly zero. Around the summit of Mt. Madison the minor folds plunge systematically southwest at low angles, with few exceptions. Between the Glen House and Dolly Copp Camp the plunge of the minor folds is toward the west and southwest.

Wherever data are available, the minor folds have essentially the same plunge as the major folds. Consequently, in those places where the larger structure is obscure, the attitude of the minor folds may aid in reconstructing the major folds.

Major folds.—Whereas the minor folds may be observed directly, the major folds are inferred from various data. As shown in Plate 9, sections BB′ and CC′, the Presidential Range is on the southeast flank of the Jefferson dome. Within the Presidential Range the strata are thrown into a series of major folds associated with which are countless minor folds. In Plate 9 the attitude of the minor folds is accurately shown, but it has been necessary to magnify the size and decrease the number because of the scale employed.

The major folds are shown in Plate 9 and even more completely in Figure 5, which is a structure-contour map of the altitude of the Fitch formation above sea level. The contour interval is 1000 feet and no attempt has been made to show the minor folds.

The axes of the major and minor folds trend north-south in the southern part of the Presidential Range, but toward the north they swing into northeasterly trends. The en echelon pattern is shown on the structure-contour map. Many of the folds can be traced for only 1, 2, or 3 miles before dying out.

Several of the major folds are asymmetric, and in all cases the southeast limb of the anticline is the steeper. These relations are shown by the anticline at Mt. Clay, the anticline northeast of Mt. Clay, the syncline

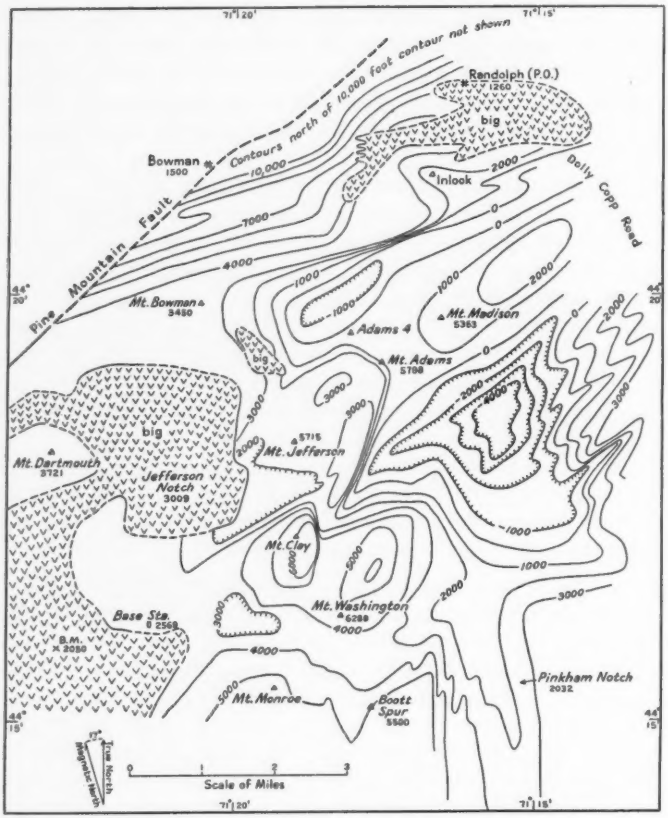


FIGURE 5.—*Structure-contour map of the Presidential Range*

Contours show altitude relative to sea level of the base of the Fitch formation. The contour interval is 1000 feet. big = Bickford granite.

4 miles northeast of Mt. Washington, and the limb 1½ miles north of Mt. Adams. Large scale overturning does not occur.

The axis of one major syncline lies half a mile west of Pinkham Notch. The attitude of the bedding on the ridge half a mile south of Cutler River is shown on Plate 10, and a structure section based on additional field data is given in Figure 4.

The east limb of this fold is also exposed on Cutler River and the headwaters of Peabody River.

The axis of a major anticline is 1½ miles west of Pinkham Notch. The east limb is the west limb of the syncline just described. The west limb of the anticline is

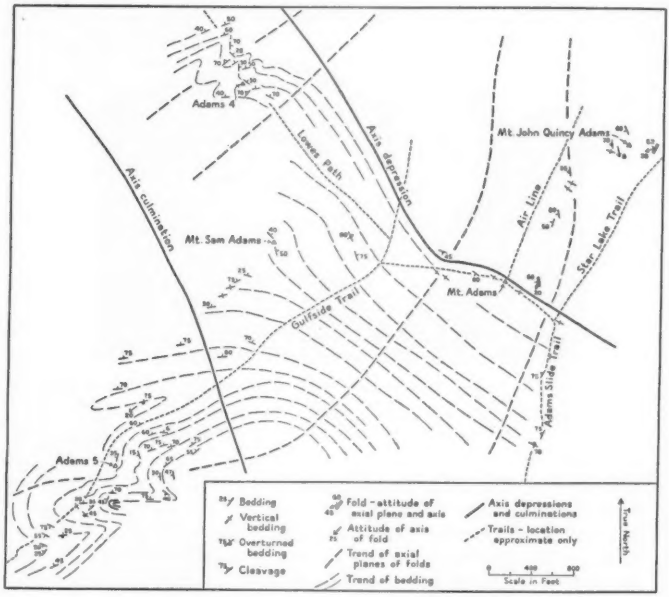


FIGURE 6.—*Sketch map of area around Mt. Adams*

The trend of the axial planes of the folds, shown by the heavy long broken lines, ranges from east to north. The cleavage parallels the axial planes of the folds and dips 35 to 70 degrees northwest. An axis culmination lies between Adams 4 and Adams 5. On Adams 4 the folds plunge northeast. On Adams 5 the folds plunge slightly south of west. On Mt. Adams the beds strike northwest and are essentially vertical. They are considered to be on the northeast nose of a fold with a vertical plunge. The folds to the northeast plunge gently south and an axis depression lies just northeast of the summit of Mt. Adams.

exposed on the north wall of Tuckerman Ravine, and on the east slope and summit of Boott Spur. This anticline plunges north and can be followed toward the north for only a mile. Further west, in the vicinity of Mt. Monroe, another anticline plunging gently north is exposed on the headwall of Oakes Gulf.

A doubly plunging anticline lies near the summit of Mt. Washington (Fig. 5). The east limb, greatly modified by minor folds, is exposed in Huntington Ravine and the west limb, accompanied by some minor folds, may be seen on the headwall of Great Gulf. The southern nose, on which the summit of Mt. Washington is located, is characterized by strata dipping to the south and minor folds plunging in the same direction (Pl. 10). The northern nose is shown in the vicinity of the Halfway House by beds dipping north and folds plunging in the same direction (Pl. 10). The anticlinal structure of Mt. Washington is also shown in Plate 9, section FF′.

Another doubly plunging anticline centers about Mt. Clay. The attitude of the steeply dipping eastern limb on the eastern slope of the mountain is shown principally by the Fitch formation. The southern nose is particularly well exposed south of the Cog Railroad (Pl. 10), where the beds strike east-west and dip south at angles

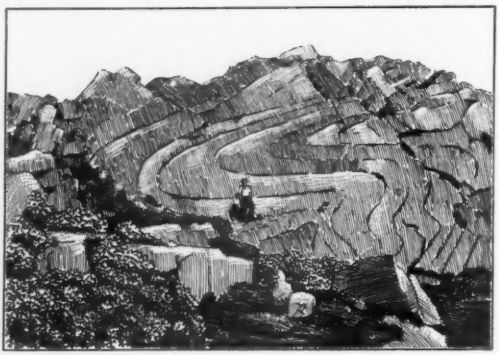


FIGURE 7.—*Fold in Littleton formation*
Near top of Mt. John Quincy Adams. The perspective makes the fold appear more recumbent than it actually is.

of 25 to 60 degrees. The strata are corrugated by a series of minor folds plunging south at the same angle as the dip of the bedding. The northerly plunge of the minor folds on the northern nose may be observed on the north slope of Mt. Clay.

An anticlinal axis lies 1 mile east of the summit of Mt. Jefferson. The average dip of the western limb, 1000 feet in a mile, is rather low (Pl. 9, section DD'), but countless minor folds are so abundant that no regional dip is apparent in the field. The minor folds on the southwest flank of this fold plunge southwest and are well displayed on the southern slopes of Mt. Jefferson.

Another anticline is located 1 mile northeast of the summit of Mt. Jefferson. Field data for this area are given in Figure 6. On Adams 4, which lies 0.6 mile northwest from Mt. Adams, the structural relations are unusually clearly displayed (Fig. 6). The strata are thrown into a series of folds which plunge 25° to 40° NE., and the axial planes dip steeply to the northwest. A well-defined cleavage, which cuts across the bedding, strikes northeast, and dips 70° NW., is essentially parallel to the axial planes of the folds.

The beds around the summit of Mt. Adams present what at first appears to be an anomalous attitude. They strike west-northwest and dip very steeply, 10 degrees either side of the vertical. Inasmuch as these beds strike directly toward those of Adams 4, they probably occupy a similar structural position on the nose of a northeasterly plunging anticline; but the plunge is vertical. Three quarters of a mile southwest of Mt. Adams it is equally apparent that the folds plunge to the west and southwest (Fig. 6) at an average of 45 degrees.

On Mt. John Quincy Adams, 0.3 mile northeast of the summit of Mt. Adams, minor folds which plunge 15 to 32 degrees south are common (Fig. 7). Similar southerly plunges are found only 400 feet east of the summit of Mt. Adams.

Mt. Madison is on the nose of a southwesterly plunging anticline. The minor folds, well exposed on the upper 300 feet of the mountain, plunge southwest (Pl. 10), although there are a few exceptional northeasterly plunges. The characteristic south-

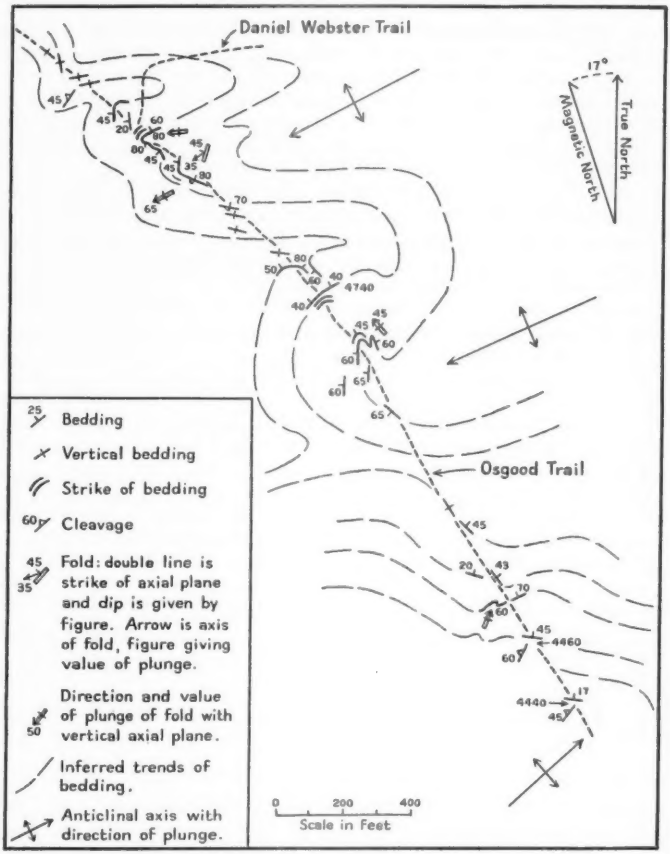


FIGURE 8.—*Structure on Osgood Ridge*

Osgood Ridge extends southeast from the top of Mt. Madison. The inferred trend of the bedding is shown by the long broken lines.

westerly plunges are found a mile northeast of the summit. The complexity of the crest of this anticline is indicated by the structure exposed on Osgood Ridge, between altitudes of 4440 and 4830 feet (Fig. 8). Considerable variation in the attitude of the bedding exists and minor folds may be observed. The cleavage, striking uniformly northeast and dipping northwest at angles of 45 to 60 degrees, is considered to be

parallel to the axial planes of the folds. At the southeast end of the ridge the folds plunge northeast, at the northwest end they plunge southwest.

A broad syncline, in which the Fitch formation is more than 4000 feet below sea level, lies 1½ miles southeast of Mt. Adams and Mt. Madison (Fig. 5). On the south side of this basin the strata dip north and northeast, at angles as high as 70 degrees, as may be seen east and west of the Halfway House (Pl. 10). One mile south of the Halfway House the plunge averages zero. Toward the north the northerly plunges become greater, attaining a maximum of 60 degrees. Even as far north as the West Branch of the Peabody River the folds plunge 38 degrees north.

Along the east and northeast sides of this basin the strata are thrown into subsidiary folds, as indicated by the outcrop pattern of the Fitch formation. The minor folds plunge southwest at angles as high as 60 degrees.

The northwest limb of a subsidiary syncline along the east side of the map (Fig. 5) is essentially vertical. This limb is particularly well exposed along the West Branch of the Peabody River. The minor folds plunge steeply southwest, and their pattern indicates that many of the beds which dip steeply northwest are overturned.

The southwestern nose of a syncline 1 mile northwest of Mt. Adams is marked by the southerly trend of the Fitch formation, 1 mile east of Mt. Bowman (Pl. 10). The minor folds in this area plunge northeast at angles ranging from 20 to 47 degrees.

The strata in the northern part of Figure 5 are on the south limb of the great dome of the Oliverian magma series. Between Mt. Bowman and Bowman a minor fold, shown on the structure-contour map, interrupts the otherwise uniform southerly dip. Further northeast a rather broad anticline, plunging northeasterly, develops and is indicated by an area of northerly strikes 2 miles east of Bowman (Pl. 10). A syncline lies between this anticline and the irregular stock of Bickford granite south of Randolph. The core of this syncline is well exposed on the Inlook (Fig. 5), where the attitude of the minor folds is highly variable. The axes strike southwest, south, southeast, east, and northeast, but the most common folds plunge southeast at angles ranging from 45 degrees to vertical.

Culminations and depressions.—Areas of plunging folds are often characterized by culminations and depressions, which trend at right angles to the axes of the folds. The folds plunge away from the culminations and toward the depressions.

In the Presidential Range there are several culminations and depressions. A depression, trending east-west, lies north of Mt. Monroe and Boott Spur, but south of Mt. Clay and Mt. Washington. South of the depression the major and minor folds plunge north, but north of the depression the plunges are south. Figure 2 of Plate 3 shows this depression.

A culmination, likewise striking east-west, passes through the highest parts of the anticlines near Mt. Washington and Mt. Clay. A depression trends northwest between Adams 4 and Mt. Madison and extends southeast to the limits of the map but dies out toward the northwest. A poorly defined culmination, which lies 1½ miles northeast of Mt. Madison, strikes west-northwest.

The major folds of the Presidential Range plunge in unison. The structurally highest area is near Mt. Washington and Mt. Clay. The

folds, with some minor oscillations, plunge northeast for 3 miles, where a structural low is reached. Still further to the northeast the folds rise again to a culmination, which, however, is not as high as that around Mt. Washington. The folds also plunge south from the highest culmination around Mt. Washington and a mile to the south reach a depression, south of which they rise again.

Schistosity and cleavage.—Schistosity is well-marked in many parts of the area, although on the upper slopes of the range massiveness is characteristic. The schistosity is due to parallel plates of biotite and muscovite, and flattened grains of other minerals, such as quartz and feldspar. In most instances the schistosity parallels the bedding. In those outcrops where such relations could not be demonstrated, the schistosity was recorded and is shown on Plate 10. In most cases it can be considered to represent the bedding.

A second more restricted structure may be referred to as cleavage. Most of its properties are those typical of fracture cleavage, and the closely spaced fractures cut across bedding. In many cases they nearly parallel the axial planes of the folds, but this relationship is not as close as for typical flow cleavage. Moreover, in many places the minerals immediately adjacent to the fracture tend to parallel the cleavage. Inasmuch as the cleavage is a shear phenomenon, it may be properly described as fracture cleavage.

Examples of cleavage at a distinct angle to the bedding and essentially parallel to the axial planes of the folds are shown by Figure 9.

The cleavage is best developed in the southeast and northeast corners of Plate 10. In the southeast it strikes north, but in the northeast it has assumed a northeasterly trend. In the southeast the cleavage dips east, but farther west it becomes westerly. In the northeast the dip is toward the northwest. The attitude of the cleavage is similar to the attitude of the axial planes of the folds.

Lineation.—Lineation is represented in the Mt. Washington area by (1) parallel elongate minerals, such as sillimanite, pseudo-andalusite, and hornblende; (2) parallel streaks of such platy minerals as biotite; and (3) "crinkles," which are small folds with an amplitude and wave length of a fraction of an inch. Elongated pebbles are absent, for the original strata were either devoid of conglomerates or metamorphism has obliterated the pebbles.

The lineation generally parallels the axes of the folds. In those areas where the minor folds are absent, the lineation may offer an important clue to the structure, although this is rarely necessary in the Mt. Wash-

ington area because minor folds are so abundant. In some cases two or even three sets of "crinkles" may be displayed on a single bedding plane. The younger sets are generally related to minor fractures and are not parallel to the axes of the folds.

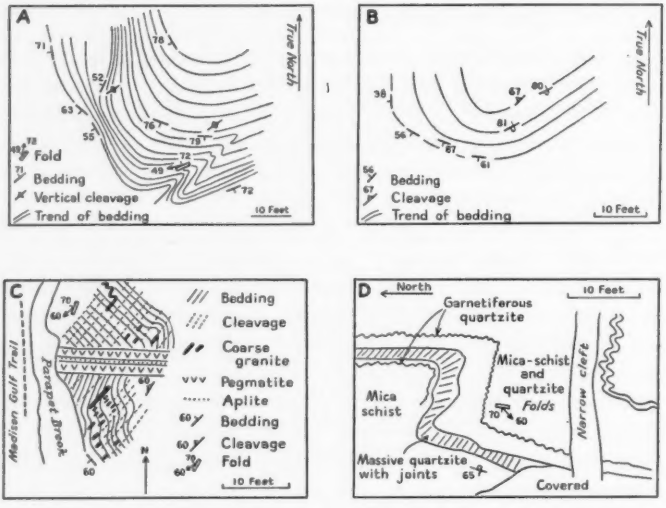


FIGURE 9.—*Sketch maps showing cleavage parallel to axial planes of folds*

A = ridge south of Culhane Brook, altitude 1600 feet; both A and B are about 1 mile south-southwest of Dolly Copp Camp. B = ridge south of Culhane Brook, altitude 1540 feet. C = Parapet Brook, altitude of 4150 feet; Parapet Brook drains the south slope of Mt. Madison. D = West Branch of Peabody River, altitude 1750 feet; 100 feet below the Osgood Ridge.

FAULTS

General statement.—Faults are recognized primarily by silicified zones. The major fault is the Pine Mountain fault, but several smaller ones have also been mapped (Pl. 1).

Pine Mountain fault.—On the northern slopes of the Dartmouth Range are a number of isolated silicified zones, where the rocks have been almost completely replaced by silica and are cut by innumerable quartz veins. The individual zones strike northeast, wherever data are available, and are aligned in a northeasterly trending zone. Elsewhere in New Hampshire silicified zones are located along faults (Billings, 1937, p. 528; C. A. Chapman, 1939, p. 165).

The most prominent silicified zone is found on Pine Mountain (Pl. 1), and the shiny white outcrops at the top of the mountain are conspicuous even from the

Presidential Range. The silicified zone, forming the summit of the mountain is bordered on the southeast by a 100-foot cliff. The zone strikes N. 40° E., is at least 80 feet wide, and crops out for several hundred feet along the strike.

There is a second exposure nearly a mile to the southwest. The silicified zone, exposed on top of a small knob, apparently strikes northeast and is 15 feet wide. A third exposure, 2.4 miles southwest of Pine Mountain, holds up a distinct terrace, and is at least 100 feet wide. A fourth exposure, 3.5 miles southwest of Pine Mountain, is a small, inconspicuous outcrop on the trail at an altitude of 2640 feet. The most southwesterly exposure, 4 miles southwest of Pine Mountain, is a silicified zone about 20 feet wide, striking N. 62° E.

The precise course of the fault northeast of Pine Mountain is uncertain. In the vicinity of Bowman there are several silicified zones. On the geological map the central one has been correlated with the Pine Mountain fault. The northern exposure, however, might have been chosen with equal justification. An isolated exposure of silicified rock 3 miles northeast of Bowman has been shown as representing the continuation of the fault.

There is, of course, important stratigraphic evidence of faulting. South of Bowman the Ammonoosuc volcanics, occupying a belt 1.3 miles wide, abut against the fault and have not been observed to the west. This type of evidence, however, is not as imposing in the field as the geological map would suggest. The section of the Ammonoosuc volcanics along the Israel River is reasonably good and complete. However, west of here exposures of the volcanics are lacking, except north of Mount Bowman, near the contact with the Partridge formation. Similarly, in the triangle bounded by the Israel River, the South Branch of the Israel River, and the fault, the Oliverian magma series has been mapped on the basis of float and doubtful exposures.

The southeast side of the fault is downthrown. Higher structural units, such as the Ammonoosuc volcanics and the gneiss of the Partridge formation, lie on the southeast side of the fault, whereas a lower structural unit, the Oliverian magma series, lies to the northwest. Nothing is known about the dip of the fault; the fact that the trace is independent of topography indicates a steep dip.

Minor faults.—The best-exposed minor fault, 2.2 miles S. 65° E. from Bowman, strikes N. 55° E., dips 65° SE., and displays a 1-foot slickensided zone. The minor irregularities on the fault plane indicate a normal fault. Gneiss of the Partridge formation on the southeast side has been dropped against Ammonoosuc volcanics and Bickford granite on the northwest side. No direct data concerning the amount of displacement are available, but it has probably been slight. No silicification has occurred along this fault.

The silicified zone 2.3 miles S. 80° W. from Bowman, at an altitude of 1460 feet on the South Branch of the Israel River, strikes N. 50° E., being 100 feet long and 20 feet wide. The principal shear planes strike N. 45° E. and dip 60° SE.

The silicified zone 1 mile N. 80° E. of Bowman is exposed in a small brook between altitudes of 1430 to 1460 feet. Fault planes within this zone strike from N. 40° E. to N. 58° E. and dip 55° SE. Striations pitch 48 degrees in a direction N. 83° E. The hanging wall has apparently gone down; the faulting is normal, and, assuming that the striations indicate the average direction of motion, there has been an easterly component to the downward movement.

Many small silicified zones have been observed in the Presidential Range. In many cases they are zones of weakness and are followed by small gulleys.

Generalizations.—Certain generalizations may be made concerning the faults and silicified zones, including the Pine Mountain fault: (1) they strike northeast, although data are lacking for two faults; (2) they dip 55° to 65° SE., in the two instances where data are available; and (3) in two cases the southeast side is downthrown, as is probably true in a third —in others data are not available. It is probable that these faults are all part of one set. From analogy with the minor faults the Pine Mountain fault is probably a normal fault dipping southeast.

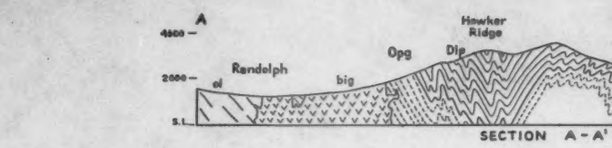
DARTMOUTH RANGE

General statement.—Study of the structure of the Dartmouth Range, which extends from Deception Brook to the South Branch of Israel River, is difficult (Pl. 1). On the northern slopes exposures are rare and this portion of the map is an approximation. Although exposures on the southern slopes, particularly in the streams, are reasonably good, the lack of a key bed handicaps structural interpretation.

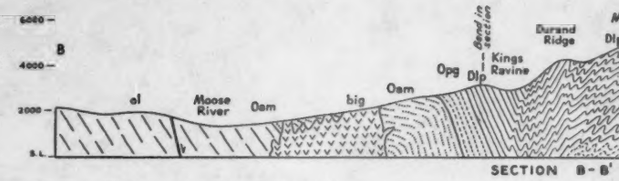
Southeast of the Pine Mountain fault.—Southeast of the Pine Mountain fault the Dartmouth Range is composed of gneiss and Bickford granite. The gneiss is the southwesterly continuation of a belt which can be traced through Mt. Bowman from the northern slopes of the Presidential Range. The gneissic structure in general strikes east-northeast, and dips range from 50° to 80° SE. The trend of the range is parallel to the strike of the gneissosity, so that on Mt. Dartmouth, where the average strike of the gneissosity is northwest, the range likewise trends northwest. Locally the gneissosity varies from the usual trend, implying some folding.

Practically no other data are available. One minor fold has been recorded near the top of Mt. Deception. It is apparent, both from the tectonic map (Pl. 10) and the structure sections (Pl. 9, section HH′), that the Dartmouth Range is on the southeast limb of a great anticlinal dome formed by the Oliverian magma series.

Presumably the gneissosity parallels the original bedding, as in the Presidential Range. If the Dartmouth Range, southeast of the Pine Mountain fault, is the limb of a major fold unmodified by minor folds, the gneiss is at least 5000 feet thick. In that case the Fitch formation should appear somewhere in the range. Moreover, the pattern of the Fitch (Pl. 1) suggests that the gneiss of the Dartmouth Range is stratigraphically beneath it. If this is so, and the Partridge formation is only 1400± feet thick, the gneiss of the Dartmouth Range must be thrown into a series of isoclinal folds, the axial planes of which dip southeast. The structure has been so shown in section HH′ of Plate 9, but the representation is strictly diagrammatic.



SECTION A-A'



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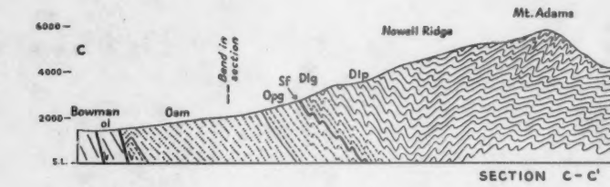

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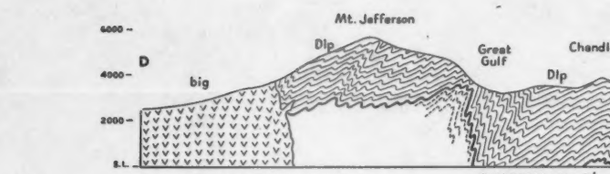

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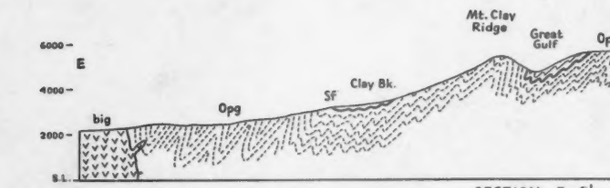

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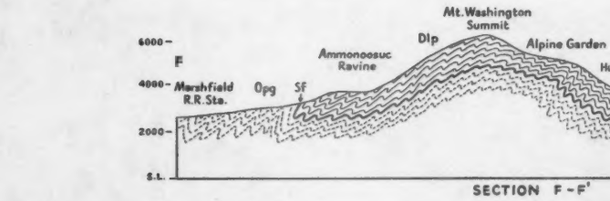

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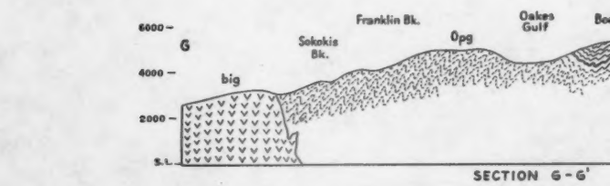

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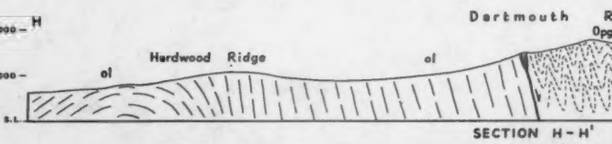
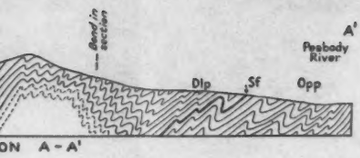


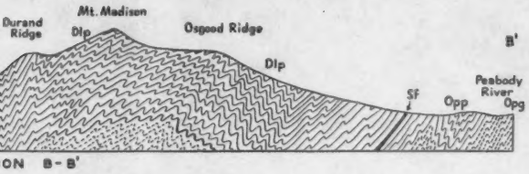
SECTION H-H'

STRUCTURE SECTION

Taken along lines shown in Plate 1. Vertical scale in feet; horizontal scale same as ve
Partridge formation; Opp = paraschists of Partridge formation; Sf = Fitch formation;
Littleton formation; ol = Oliverian magma series; big = Bickford granite; t = tuff and



ON A - A'



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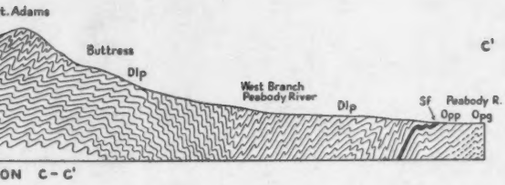

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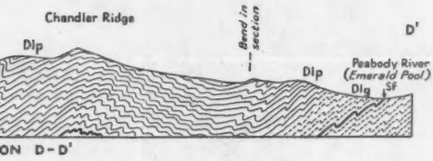

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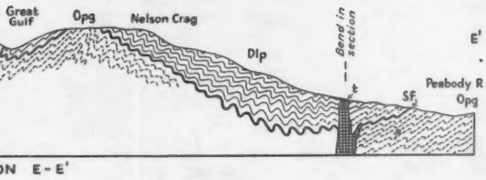

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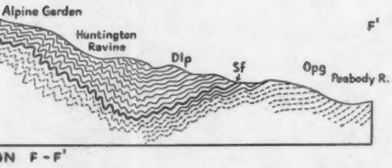

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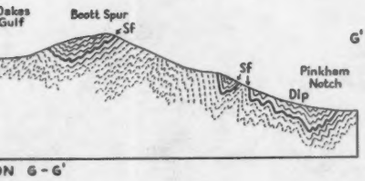

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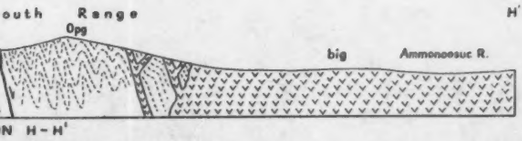

N H - H'

SECTIONS

e same as vertical scale. Oam = Ammonoosuc volcanics; Opg = gneiss of
n formation; Dlg = gneiss of Littleton formation; Dlp = paraschists of
= tuff and breccia of volcanic vents.

Northwest of the Pine Mountain fault.—Data are scanty on the northwestern slopes of the Dartmouth Range, especially northwest of the Pine Mountain fault. On the northwestern slope of Mt. Deception the gneiss is exposed at altitudes as low as 2600 feet. The dip of the gneissosity ranges from 40° to 70° SE., and the rocks are probably thrown into a series of isoclinal folds trending east-northeast and overturned toward the northwest. Whether there is a belt of Ammonoosuc volcanics between the gneiss of the Partridge formation and the granitoid rocks of the Oliverian magma series is questionable. The critical area without exposures is 2000 feet wide, and this is the maximum possible width of such a belt. The erratics and float, however, indicate a narrow belt of Ammonoosuc volcanics as shown on the geological map. Apparently the rocks of the Oliverian magma series have injected higher stratigraphic units here than east of Bowman, and the Ammonoosuc volcanics have been nearly entirely eliminated.

BICKFORD GRANITE

The Bickford granite occurs in several stocks. A large one envelops the Dartmouth Range on the west, south, and east. One small stock is in the Ravine of the Castles, in the headwaters of the Israel River. A stock of intermediate size lies south of Randolph. Another is located near Dolly Copp Camp. The older rocks surrounding these stocks are commonly injected by numerous dikes and sills of the Bickford granite. The granite of the stocks is typically massive, suggesting that it is younger than the major orogenic period. Moreover, undeformed dikes of this rock cut the folded paraschists and gneisses.

Although the Bickford granite is essentially massive, a planar structure was observed in a few localities (Pl. 10). This structure is due to the parallelism of feldspar crystals or less commonly to parallel plates of biotite. Reference to the tectonic map reveals that this planar structure in general strikes parallel to the regional trend lines, suggesting that it resulted from compressive stresses in operation during the waning stages of the orogeny.

The mechanics of intrusion of the Bickford granite, particularly of the largest stock, deserves consideration. In several places the contact is discordant (Pl. 10), as, for example, 1 mile south of Jefferson Notch, 1 mile south of Mount Deception, and 1 mile southeast of Pine Mountain. Concordant relations exist only where the contact parallels the regional trend of the structures. The discordant relations suggest stoping. The manner in which the Bickford granite has shattered the surrounding rocks indicates that the country rock was relatively brittle at the time of intrusion. All the conditions were favorable for the stoping of the country

rock. Xenoliths have been observed in the Bickford granite but are not especially abundant.

On the other hand, reference to either the geological map (Pl. 1) or the tectonic map (Pl. 10), shows that the trend lines in the country rock wrap around the major stock of Bickford granite. Northwest of the stock the structures trend east-northeast; east of the stock the structures trend nearly north-south. Such relations suggest that the Bickford granite inserted itself like a great wedge, shoving apart the older rocks.

It is concluded that the Bickford granite forced its way into the older brittle rocks. The rocks surrounding the rising magma were mechanically shattered and broken, and countless dikes and sills of magma forced into the fractures. Blocks of country rock, engulfed by the magma, were assimilated, the extent of the assimilation depending upon whether the blocks were incorporated in the early or late stages of the rise of the magma.

The relative importance of the two major processes involved cannot be evaluated at present. Additional work in the Mt. Washington area might shed more light on the subject, but it is felt that most of the critical data available have been obtained. Study in the surrounding areas may aid in attempts to reach a more definite conclusion.

AGE OF THE STRUCTURAL FEATURES

The main orogenic period is considered to have been Acadian (middle or late Devonian). Two methods of attack lead to this conclusion.

The deformation is obviously younger than the metamorphosed sediments involved, which are Ordovician (?), Silurian, and lower Devonian. In the Franconia quadrangle, the Moat volcanics rest unconformably on the paraschists of the Littleton formation (Williams and Billings, 1938, p. 1025). Sufficient time elapsed between the deposition of the lower Devonian Littleton formation and the eruption of the Moat volcanics for folding, metamorphism, intrusion of the Oliverian and New Hampshire magma series, and deep erosion. Hence it is unlikely that the Moat volcanics are older than Mississippian. They are the extrusive phase of the White Mountain magma series, which is consanguineous with the pre-Pennsylvanian Quincy-Blue Hill group of the Boston region (LaForge, 1932, p. 37). Therefore, the Moat volcanics cannot be younger than Mississippian, and the orogeny is younger than the lower Devonian, but older than the Mississippian.

A second fact bearing on the age of the folding relates to the Littleton-Moosilauke area where evidence shows that the Bethlehem gneiss was syntectonic (Billings, 1937, p. 537). Subsequently Shaub (1938) demon-

strated that some of the pegmatites derived from the Bethlehem gneiss in the Cardigan quadrangle contained uraninite of Devonian age.

Evidence has been offered elsewhere (Billings, 1937, p. 535-536) that the intrusive domes of the Oliverian magma series of the Littleton-Moosilauke and Mascoma areas preceded the folding but are younger than the lower Devonian. Hence they are considered to be early Acadian.

The Bickford granite, which belongs to the New Hampshire magma series, is younger than the main deformation, but it apparently felt the waning effects of the compressive stresses.

The Pine Mountain fault is younger than the Ammonoosuc volcanics, the Partridge formation, the Oliverian magma series, and the Bickford granite, as all these rocks are cut by the fault. There is no evidence that the fault cuts the Conway granite or displaces the contact of the Conway granite. Although such evidence is not very significant because of the poor exposures, it is probable that the fault is older than the Conway granite. If so, the fault is late Devonian or early Mississippian.

METAMORPHISM

GENERAL STATEMENT

The original sedimentary rocks of the area have been thoroughly recrystallized and metamorphosed into a variety of crystalline schists. The rocks involved in these transformations were originally shale, sandstone, dolomitic sandstone, calcareous shale, basaltic tuff, quartz latite tuff, and rhyolite tuff. All of the metamorphic rocks, with the possible exception of those in a small area east of Mt. Madison, lie in the high-grade zone of metamorphism. However, there are several subzones.

The minerals in the paraschists of the Littleton and Partridge formations did not form simultaneously, and a definite paragenesis may be established. The problem of chemical changes during the metamorphism is given special consideration. To what extent have new elements been added to the rocks and how much material has been taken away by moving solutions? In particular, what is the origin of the gneiss in the Partridge formation and at the base of the Littleton formation? It is generally recognized that regional metamorphism of the type here considered is associated with a period of orogeny. In the Mt. Washington area this is true, and the stages in the recrystallization can be correlated with the orogenic stages.

ZONING AND SUBZONES IN THE PARASCHISTS

All rocks in the area, with the exception of one small region, lie in the high-grade zone of metamorphism, corresponding to the sillimanite zone of the British (Harker, 1939) and the katazone of Grubenmann and

Niggli (1924). The index mineral of the high-grade zone, sillimanite, is generally present throughout the Presidential Range in the paraschists and gneisses. Moreover, the groundmass of the rocks is much coarser grained than in typical middle-grade schists.

No direct criteria determine whether the actinolite and diopside granulites of the Fitch formation belong in the high-grade or middle-grade zone. As this unit is thin and is overlain and underlain by typical high-grade schists, it must be high-grade. Similarly, the Ammonoosuc volcanics, being nonaluminous, carry no index minerals for distinguishing between middle-grade and high-grade metamorphism. On the headwaters of Moose River, at an altitude of 1625 feet, and well within the area of Ammonoosuc volcanics, mica schist containing sillimanite is interbedded with the typical volcanics, indicating that the Ammonoosuc volcanics are high-grade.

In the description of the paraschists of the Littleton and Partridge formations it has been pointed out that original argillaceous sediments have recrystallized to rocks which differ greatly in their physical appearance and somewhat in their mineral composition. For descriptive purposes these rocks have been classified as: (1) coarse andalusite schists; (2) coarse pseudo-andalusite schist; (3) coarse rough pseudo-andalusite schist; (4) intermediate pseudo-andalusite schist; (5) shiny fine-grained pseudo-andalusite schist; (6) staurolite schist; and (7) medium-grained massive schist. In Figure 3 two varieties in several cases have been grouped under one symbol.

On the summits and upper slopes of the Presidential Range the original argillaceous sediments have been converted to coarse rough pseudo-andalusite schists. Lower down on the slopes, the rocks are not as coarse and are designated as intermediate pseudo-andalusite schist.

The coarse andalusite schists and the coarse pseudo-andalusite schists occupy an area extending from half a mile west of Pinkham Notch to a point 2½ miles to the north. The coarse andalusite schists are not extensive, because in most areas the andalusite has been completely altered to sericite or muscovite, forming the coarse pseudo-andalusite schists. Where the andalusite is partly altered the rocks constitute a transitional variety.

On the west side of the valley at Pinkham Notch the original argillaceous sediments are now shiny pseudo-andalusite schists, traceable north-northeast for 4½ miles. In the vicinity of the Mt. Washington Auto Road and the Glen House are local areas of medium-grained massive schist.

The staurolite schists occupy slightly more than a square mile southeast of Mount Madison. Although the coarse rough pseudo-andalusite schist and the intermediate pseudo-andalusite schist carry staurolite, they

also carry the index mineral sillimanite and are relatively coarse. The staurolite schists have no sillimanite and the matrix is fine-grained. Such being the case, these rocks should be considered middle-grade.

The distribution of the different varieties is not due to any difference in stratigraphic position, as is quite apparent from Figure 3. In the vicinity of Mt. Clay coarse, rough pseudo-andalusite schist directly overlies the Fitch formation. Around Mt. Monroe and Boott Spur the intermediate pseudo-andalusite schist is directly above the Fitch formation. On the Cutler River, however, shiny pseudo-andalusite schist lies on the Fitch. Near summit 2587, 2 miles east-southeast of Mt. Madison, staurolite schist overlies the Fitch formation. Thus the varieties are not due to stratigraphic position but to differences in physical conditions during metamorphism. Variations in the character of moving solutions may have played a role. What these differences were is discussed in a later section.

PARAGENESIS IN THE PARASCHISTS

Sequence.—In the paraschists of the Littleton formation, particularly the coarser varieties, it is apparent that the minerals are not contemporaneous. The general sequence is shown in Fig. 10. At least three major stages may be recognized: andalusite belongs to the first; sillimanite, staurolite, garnet, tourmaline, and much of the muscovite and biotite belong to a second; and chlorite and sericite belong to the third stage. Some muscovite and biotite may belong to the first stage, although direct evidence is lacking. Quartz presumably belongs to both the first and second stages.

Evidence.—The evidence on which the paragenesis has been determined is comparatively simple and clear. Andalusite is older than muscovite, sillimanite, and staurolite; in many instances andalusite is surrounded by a pseudomorphic shell of one or more of these minerals. Staurolite and muscovite are considered to outlast sillimanite. There is no evidence for this in the Mt. Washington area, but in the Littleton-Moosilauke area shells of staurolite and muscovite surrounding sillimanite are common (Billings, 1937, p. 551). Muscovite, however, outlasted staurolite, for pseudomorphic shells of muscovite around staurolite are common in the staurolite schists.

Garnet is apparently older than staurolite. In many specimens euhedral crystals of staurolite are associated with irregular corroded grains of garnet, indicating that garnet was out of equilibrium and was dissolving when staurolite was forming. Biotite is believed to be essentially contemporaneous with muscovite, chiefly because it shows essentially the same amount of deformation. Tourmaline is considered to

be contemporaneous with staurolite as it likewise occurs in euhedral grains.

Sericite and chlorite belong to the third stage. Much of the andalusite is completely sericitized. Moreover, the staurolite has thin pseudomorphic

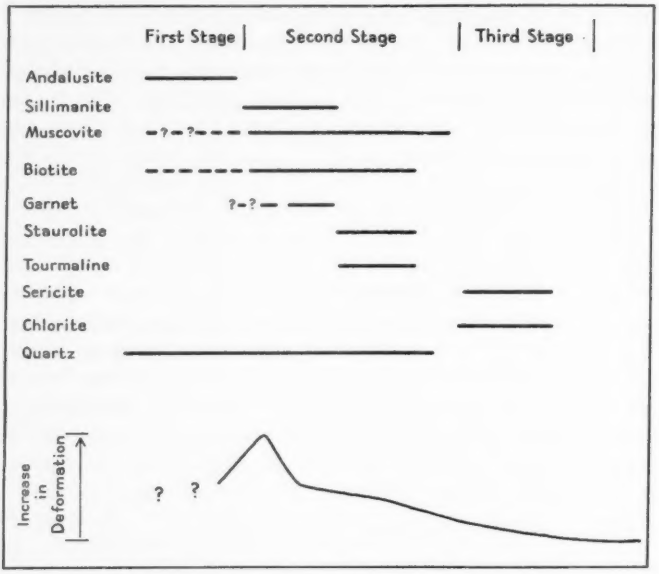

FIGURE 10.—*Paragenesis of metamorphic minerals*
Refers primarily to the coarse rough pseudo-andalusite schist of the Littleton formation. The abscissa represents time, the younger minerals appearing to the right.

shells of sericite, and in many specimens a sericitized andalusite is surrounded by a muscovite shell. This implies that the outer part of an andalusite crystal was replaced by muscovite and subsequently that part of the andalusite not already altered was converted to sericite. Chlorite is younger than biotite and garnet, as both of these minerals are extensively chloritized. Chlorite and sericite are associated along planes of fracture cleavage and are apparently contemporaneous.

CORRELATION OF PARAGENESIS AND DEFORMATION IN THE PARASCHISTS

Problem.—The chronological relation between paragenesis and deformation is an important problem. In general the recrystallization was syntectonic, but some minerals are older than the culmination of the orogeny, others are younger. Conceivably the deformation was not simultaneous

throughout the region, and the same may be true of the recrystallization. Consequently some rather complicated relations may occur.

Coarse rough pseudo-andalusite schist.—In these rocks many of the andalusite pseudomorphs have been folded with the strata (Pl. 7, fig. 4). In some thin sections the sillimanite which partially replaces andalusite is folded and broken. In such cases it is evident that the minor folds are younger than the andalusite and sillimanite. There is some evidence, to be discussed later, that the large folds had already developed when andalusite and sillimanite formed.

The muscovite generally occurs as large undeformed plates, but locally it is broken by sharp flexures. This indicates that some deformation outlasted the recrystallization, but it is distinctly a subordinate phenomenon. The biotite shows relations similar to muscovite. Staurolite and tourmaline are in euhedral grains which lie in diverse orientations. They are considered, therefore, to be relatively late minerals.

Intermediate pseudo-andalusite schist and shiny pseudo-andalusite schist.—In these rocks the pseudomorphs after andalusite show a distinct lineation in many localities. Likewise the biotite in the shiny pseudo-andalusite schists may show a linear parallelism. The nature of the lineation indicates that it is syntectonic, not mimetic. In these rocks, therefore, the andalusite and biotite, and probably other minerals, are contemporaneous with the deformation. Little crinkles, however, which are due to small folds in the schistosity, parallel the lineation shown by the elongation of the minerals. These crinkles indicate that deformation outlasted recrystallization.

Fitch formation.—Many specimens of this unit show an excellent lineation. The actinolite in the granulites is in stubby grains which lie parallel to one another, indicating that they formed contemporaneously with the deformation.

Conclusions.—The recrystallization of the Fitch formation and the shiny pseudo-andalusite schists was contemporaneous with the formation of the minor folds. In the coarse rough pseudo-andalusite schist, however, the recrystallization largely outlasted the development of the minor folds.

METAMORPHISM, STRUCTURE, AND TOPOGRAPHY

A comparison of the structure-contour map (Fig. 5) with the map of the metamorphic subzones (Fig. 3) shows that there is a rather close correlation between the anticlines with closure and the coarsest schists. The coarse rough pseudo-andalusite schist and the intermediate pseudo-andalusite schist are characteristically developed on the anticlines on

Mt. Washington, Mt. Clay, Mt. Jefferson, Mt. Adams, and Mt. Madison. Conversely the fine-grained schists, such as the shiny fine-grained pseudo-andalusite schist, medium-grained massive schist, and staurolite schist, are characteristically found in the synclines and on the flanks of folds. Schists of this type are characteristic of the syncline west of Pinkham Notch—particularly the east limb—and the large syncline 3 miles northeast of Mt. Washington.

Why there should be such a structural control of the metamorphism is problematical. A possible explanation is that the development of large crystals—particularly andalusite—was favored by hot aqueous solutions. Such solutions would tend to be concentrated and trapped on the crest of anticlines with closure.

There is an interesting correlation between metamorphism, structure, and topography. The height of the Presidential Range is due to the exceedingly tough, coarse-grained schists of which it is composed. Such rocks are rare in New Hampshire, but wherever similar schists are found in the state they hold up relatively high mountains—Moosilauke, Kearsarge, Monadnock. In the Mt. Washington area, at least, the nature of the metamorphism has been controlled by the structure. Consequently the topography is closely related to the structure and the individual mountains are anticlinal. The anticlinal structure alone, however, would not suffice to cause the location of mountains. Rather, the relation is an indirect one, for the topography is controlled by the character of the metamorphism, which in turn has been controlled by the structure.

CAUSES OF SUBZONES IN THE PARASCHISTS

The subzones of metamorphism of the Mt. Washington area result from variations in the metamorphic history in different sections. There is no evidence that the subzones were buried at different depths. The temperature, however, may well have differed from place to place. Moreover, the time of maximum temperature and time of maximum deformation may have been synchronous in some localities, but elsewhere either one might have preceded the other. In the folding of strata, moreover, strong differential movements are often confined to special zones, in which highly schistose rocks may be expected.

Those rocks with large andalusite and large pseudo-andalusite crystals went through a stage which did not affect the other rocks. It was essentially a contact metamorphism, but the causative igneous body is not known.

In a later phase of the metamorphic history temperature and stress conditions favored the formation of sillimanite throughout most of the area, and this mineral either partially replaced andalusite or formed about

new centers of crystallization. However, in one area now occupied by the staurolite schists (Fig. 3) the physical conditions did not permit the formation of sillimanite.

With falling temperature, accompanied by decreasing intensity of deformation, conditions favored the formation of staurolite throughout the region. In those areas where coarse andalusite or sillimanite were present, the staurolite, along with muscovite, formed as small shells replacing the older aluminous minerals. Where these minerals had not previously formed, the staurolite developed as independent crystals.

A massive rather than a schistose structure may be due to one of two reasons. The rock may have formed in a region which was not affected by strong differential movements; or the recrystallization may have followed the deformation. The medium-grained massive schists are believed to have developed in preference to the shiny fine-grained pseudo-andalusite schists because of lack of strong differential movements.

The subzones are due, therefore, to a complicated interplay of the various factors involved in the metamorphism.

CHANGES IN CHEMICAL COMPOSITION

General statement.—One of the primary purposes of the present investigation was to ascertain the extent of chemical changes during metamorphism. This is always difficult, as primary differences may not be readily distinguished from changes due to metamorphism. Moreover, there is the danger of unconsciously selecting material with which to prove a preconceived idea.

Fitch formation.—The abundance of minerals rich in lime and magnesia, such as actinolite, diopside, and calcic plagioclase, make it clear that many of the rocks in the Fifth formation have been derived from arenaceous and argillaceous dolomites or limestones. Marble has not been observed, and even calcite is rare. There are two possible ways to explain the origin of these rocks. They may have been arenaceous and argillaceous dolomites, and the metamorphism may have taken place with little change in chemical composition, except the loss of CO_2 and perhaps H_2O; or the initial rocks may have been arenaceous and argillaceous limestone, and much of the iron and magnesia may have been introduced from extraneous sources. The first hypothesis is more likely correct, although direct proof is lacking. In the Littleton-Moosilauke area it was possible to demonstrate that the actinolite and diopside granulites were derived from arenaceous and argillaceous dolomites, for granulites are no commoner in the higher zones than dolomites in the low-grade zone (Billings, 1937, p. 547). Moreover, in the Littleton-Moosilauke area, thin beds of marble are interbedded with the granulites,

and the contacts are sharp. These relations would not be consistent with an explanation involving extensive metasomatism.

Ammonoosuc volcanics.—In the Littleton-Moosilauke area it was believed that the Ammonoosuc volcanics were metamorphosed without important compositional changes. The same is true in the Mt. Washington area.

Amphibolites are rare, but a chemical analysis was made of one specimen W56 (Table 10, column 11). The composition is similar to typical basalt (Table 10, column 13). It is a thin layer interbedded with the biotite gneisses and must be a bed of tuff.

The biotite gneisses are abundant, but the ratio of plagioclase to orthoclase varies considerably. Some, like the one listed in Table 1, column 4, do not differ much in chemical composition from some of the soda-rhyolites of the low-grade metamorphic zone of the Littleton-Moosilauke area (Billings, 1937, Table 19, column 6; Table 3, line 2). Others, such as the one listed in Table 1, column 5, compare favorably in their chemical composition with those listed in Table 3, line 3, of the Littleton-Moosilauke paper. Thus, many of the biotite gneisses of the Mt. Washington area have apparently been derived from the original volcanics without much chemical change.

Volcanics with a distinct dominance of potash feldspar, such as those listed in Table 1, columns 3 and 3a, and Table 10, column 14, have not been observed in the low-grade zone. It might be proposed, therefore, that such rocks indicate the introduction of potash. It is equally possible, however, that such rocks represent original rhyolites, which are absent from the low-grade zone or were not observed during the field work.

Paraschists.—In the high-grade zone of the Franconia quadrangle many schists derived from argillaceous sediments have been enriched in potash (Billings, 1938). Pseudomorphic shells of muscovite surrounding sillimanite constitute the field evidence, and chemical analyses supported this interpretation.

Two analyses are available for the low-grade slates from the Littleton area. Specimen W219 (Table 10, column 1) is the less aluminous type and does not differ significantly from an analysis of the shiny pseudo-andalusite schist, Specimen W229 (Table 10, column 4). The schist is slightly richer in silica and poorer in total iron and magnesia, suggesting a slightly more arenaceous original rock. However, potash is identical in the two rocks, and soda is actually somewhat less in the schist. Apparently alkalies have not been added to this particular rock during the metamorphism.

A second analysis of slate, L489, is a more aluminous variety (Table 10, column 2). Compared to the coarse rough pseudo-andalusite schists, it is somewhat richer in silica, but poorer in alumina and potash. The field evidence suggests that some potash has been introduced into the schist, for part of the muscovite is pseudomorphous after large andalusite crystals, now completely replaced by muscovite, staurolite, and sillimanite. The analysis is consistent with this hypothesis, as the coarse rough pseudo-andalusite schist contains 0.68 per cent more potash than the slate.

The shells of muscovite surrounding andalusite in the coarse andalusite schist and the coarse rough pseudo-andalusite schist and enveloping staurolite in the staurolite schist, indicate addition of potash at a relatively late stage in the paragenesis.

Much of the andalusite has been replaced by sericite. Some specimens of pseudo-andalusite have an outer shell of muscovite and a core of sericite. Moreover, euhedral staurolite crystals, which are rather late in the paragenesis, have narrow shells of sericite, indicating that the potash in the sericite was introduced rather late in the paragenesis. In any case, a tremendous difference in the size of the muscovite and the sericite indicate a distinct difference in age. The relations with staurolite indicate that the sericite is younger.

In order to see how much change took place in the metamorphism of impure sandstone to quartzite, two analyses were prepared. One, W220 (Table 10, column 8), is a low-grade sandstone from the Littleton area; the second, W222 (Table 10, column 9), is a quartzite from the Mt. Washington area. Systematic differences exist between the analyses. The sandstone is richer in silica but poorer in the other important oxides. Contrasts between the analyses could be due to differences in the original sediments, the sandstone having had more quartz and less clay than the rock from which the quartzite was derived. This interpretation, rather than ascribing the differences to changes during metamorphism, is preferred because of the evidence offered by the graphic representation described on a later page.

Gneiss.—The gneisses so common in the Partridge formation consist, over large areas, of alternating dark and light streaks in about equal proportions. The dark streaks are primarily biotite, with minor amounts of quartz, plagioclase, and muscovite (Table 2). The light layers are primarily quartz, plagioclase, and muscovite, with subordinate amounts of biotite.

After studying these gneisses in the adjacent North Conway quadrangle (Fig. 1) the writer concluded that they were injection gneisses (Billings, 1928, p. 85-86).

"Juices from the invading Chatham granite soaked through the folding sediments, greatly augmented the tendency to rock flowage, and caused recrystallization. As the granite moved up into a given horizon of the metamorphosed zone it first intruded the schists in lit-par-lit fashion and then gradually ripped them to pieces, strewing the shreds throughout the granite."

The dark portions of the gneiss were considered to be remnants of the older schists, modified by moving solutions, whereas the light-colored portions of the gneiss were believed to have been injected as thin sheets of magma.

During the present investigation a chemical analysis has been made of a gneiss composed of about equal parts of light and dark constituents (Pl. 5, fig. 1; specimen W215, Table 10, column 10; Table 2, column 1a). Such a specimen is typical of much of the gneiss (Fig. 3). The chemical composition is essentially the same as that of some of the slate (W219, Table 10, column 1), suggesting that slate has been transformed *in situ* into gneiss with little change in chemical composition. The gneiss is a little richer in soda and potash than the slate. This may mean the introduction of some alkalies, but the differences are so slight that too much significance should not be attached to them. As the metamorphic processes increased in intensity the light and dark constituents segregated into alternating bands. Turner (1940) has recently discussed this process in an admirable paper. In general the streaks followed the original bedding. A good example, however, of the light constituents concentrating along the cleavage is shown by Figure 9.

Such an explanation is inadequate for predominately light-colored gneisses (Fig. 3). The gneisses in which the dark and light constituents are equal contain only 17 per cent plagioclase, 1.71 per cent soda, and 0.59 per cent lime. The average gneiss, however, contains 36 per cent plagioclase, about 3.7 per cent soda, and about 1.4 per cent lime. Some of the light-colored gneisses contain up to 50 per cent plagioclase, and a correspondingly greater amount of soda and lime. Apparently as much as 4 or 5 per cent of soda, potash, and lime have been introduced into the light-colored gneisses.

Graphic representation.—Many devices have been employed to represent graphically the chemical relations between rocks. One of the most common systems involves the use of triangular diagrams which are unsatisfactory, inasmuch as only a limited number of the essential oxides can be incorporated. Four components can be shown on a tetrahedron, but the perspective may be confusing and even misleading.

In Figure 11, applying only to original shales or sandstones, the various oxides have been plotted against silica as the abscissa, as in the ordinary variation diagram for igneous rocks. At first this method may seem

arbitrary and unjustified for rocks of sedimentary origin. For shales and sandstones derived from thoroughly weathered older rocks, however, this method might be satisfactory. Such rocks are composed essentially of two parts, a clay portion and grains of quartz. Inasmuch as the clay

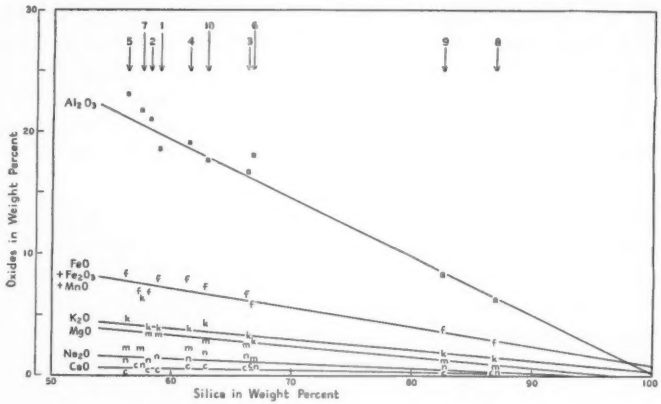

FIGURE 11.—*Graphic representation of chemical changes during metamorphism*

a = alumina; f = FeO + Fe_2O_3 + MnO, (Fe_2O_3 recalculated to FeO); k = K_2O; m = MgO; n = Na_2O; c= CaO. Numbers are the same as in Table 10. Low-grade rocks: 1 and 2 = slate. Middle-grade rock: 3 = staurolite schist. High-grade rocks: 4 = shiny fine-grained pseudo-andalusite schist; 5 = coarse rough pseudo-andalusite schist; 6 = medium-grained massive schist; 7 = muscovitized schist; 8 = sandstone; 9 = quartzite; 10 = gneiss.

portion would presumably have a uniform composition, all the other oxides would vary inversely to the silica. The other oxides should intersect the base of the plot at 100 per cent silica.

Figure 11 dispels any doubts of the theoretical justification of the method. The curves are remarkably straight and the various points are relatively close to the appropriate curve,—much better than in most variation diagrams of igneous rocks.

The curves (Figure 11) have been constructed on the analyses from the low-grade zone, for which two analyses of slate and one of sandstone are available. Although these few points may seem a rather weak basis for statistical comparisons, the fact that the points for rocks from the high-grade zone lie very close to these curves justifies the principle involved. Moreover, all the curves intersect the base close to 100 per cent silica.

In comparing the high-grade with the low-grade metamorphic rocks, any oxide which has changed notably will lie off the curves. In analysis 7

of Figure 11, potash lies several per cent above the curve. This is a muscovitized schist from the Franconia quadrangle (Billings, 1938a), where field evidence indicates addition of potash. In analysis 5, the coarse rough pseudo-andalusite schist, potash is somewhat above the curves, suggesting the possible addition of some potash. Although the amount is slight and might be within the limits of error of the method, the field data—pseudomorphs of muscovite after andalusite—indicate introduction of potash. Similarly, some potash seems to have been introduced into the gneiss (analysis 10). There is no evidence for an increase in the content of potash in the shiny fine-grained pseudo-andalusite schist (analysis 4), the quartzite (analysis 9), or the sillimanite schist from the Franconia quadrangle (analysis 6).

Soda seems to show no systematic variation except for an increase of less than 1 per cent in the gneiss. Iron seems to have increased, but in this case the left end of the curve seems to be a little too low, because the curve, as drawn through the control points, strikes the base of the plot to the right of 100 per cent silica. On an adjusted curve iron does not show any systematic variation. Magnesia, however, shows a systematic loss, amounting to over 1 per cent in some cases. Alumina seems to show an increase. However, it is more likely that analysis 1 is abnormally low in alumina, so that if the curve were based on analysis 2 alone, this oxide would show no systematic change.

Summary of chemical changes.—Changes in chemical composition during the metamorphism have been investigated by field, microscopic, and chemical methods. In general the changes in the Ammonoosuc and Fitch formations have not been great. Even in the Partridge and Littleton formations, as shown by Figure 11, many rocks seem to have undergone little or no change; this is notably true of the shiny fine-grained pseudo-andalusite schist (analysis 4), massive medium-grained schist (equivalent to analysis 6), and quartzite (analysis 9).

Potash has been added to some schists, such as the coarse rough pseudo-andalusite schist (analysis 5), but not to the same extent as the muscovitized schist of the Franconia quadrangle (analysis 7). Less than 1 per cent each of soda and potash have been added to the gneiss (analysis 10), but 3 per cent of soda and 1 per cent lime and 1 per cent potash have been added to some of the light-colored gneiss for which chemical analyses are not available.

A systematic loss of magnesia has occurred. The relative abundance of tourmaline suggests the addition of boron, and the conversion of such minerals as andalusite to muscovite indicates the addition of water.

Mechanics of the chemical changes.—The changes in chemical composition described in preceding sections were accomplished by moving solutions which added and removed material. In some rocks no changes are apparent; in others, where the changes were slight, pseudomorphs of muscovite and sericite formed after aluminous minerals. In regions of even greater alterations the introduced materials participated in the selective fusion or solution of the rocks. The solutions so generated gathered into irregular light-colored streaks, which, although they constitute 50 per cent of the rock, include less than 1 per cent of added material. In the areas of most intense alteration, where 4 or 5 per cent of soda, lime, and potash were introduced, the whole rock became a molten mush similar to magma.

The gneisses resulting from the processes described above would be called migmatites by many (Tyrrell, 1930, p. 331). However, the writer has intentionally refrained from the use of this term because it is used differently by various investigators. The process itself is similar to anatexis, metatexis, palingenesis, and granitization (Barth, Correns, and Eskola, 1939, p. 226). These terms also have been used with different meanings by geologists.

The gneisses of the Mt. Washington area formed by processes identical to those postulated for central Massachusetts by Larsen and Morris (1933). The process is similar to that described by Currier (1937) in the Chelmsford-Westford district of Massachusetts and at Milford, New Hampshire. Similar interpretations have been given to granitic rocks in other parts of North America during the last few years. Goodspeed (1939) has discussed this subject and gives an excellent list of references.

Within the gneiss small bodies of typical granitoid rocks, a few feet to several hundred feet across, have relatively sharp contacts with the gneiss and locally contain angular inclusions of gneiss. Such granitoid rocks apparently consolidated from magma injected into the gneiss. Probably these magmas were derived from deep-seated sources and are apophyses of a larger mass below.

In principle, a clear distinction should be made between gneiss which has originated *in situ* and granitoid rocks which have consolidated from magma derived from below; in practice the distinction may be difficult. Where exposures are poor or the observer is not aware of the situation, the gneiss may appear to grade imperceptibly into the granitoid rocks. Where exposures are good, however, a relatively sharp discontinuity may be recognized.

On the accompanying maps (Pl. 1; Fig. 3) the areas shown as gneiss of the Partridge and Littleton formations contain rocks of two different origins. The most abundant are gneisses derived from shales by selective

solution accompanied by the introduction of alkalies and lime. Less common, but equally important rocks occupy bodies which vary from a few feet to several hundred feet across, too small to show on the scale employed. These rocks consolidated from magma of distant origin.

SUMMARY OF GEOLOGIC HISTORY

GEOSYNCLINAL PHASE

The geological record begins with the accumulation of the Ammonoosuc volcanics, probably during late Ordovician. Not less than 5000 feet of pyroclastics, chiefly rhyolite, quartz latite, and dacite, but including some andesite and basalt, were deposited, accompanied by minor amounts of sedimentary material. The volcanic cycle was followed by the deposition of 1400 feet of sandstone and shale to form the Partridge formation. Evidence elsewhere in New Hampshire indicates a mild period of orogeny near the end of the Ordovician, accompanied by uplift and followed by erosion.

In a middle Silurian sea, which moved in from the southwest, the dolomitic sandstones and shales of the Fitch formation, not over 200 feet thick, were deposited. Late Silurian history is obscure, but during early Devonian 4000 feet or more of sandstone and shale were deposited to form the Littleton formation.

OROGENIC PHASE

Deformation.—During middle or late Devonian the strata were caught in a major orogenic paroxysm. Even before the deformation, or at least in its early stages, successive injections of the Oliverian magma series formed a large intrusive dome. This northeasterly-trending dome is in the northwestern part of the area and was inserted beneath or within the Ammonoosuc volcanics. The Ordovician (?), Silurian, and Devonian strata of the southwestern part of the area were thrown into a series of doubly plunging, en echelon folds, the axes of which trend north and northeast. Associated with the major folds are countless minor folds, the amplitudes and wave lengths of which are measured in feet, tens of feet, and scores of feet. Schistosity, parallel to the bedding, formed during the earlier stages of the orogeny. A lineation, well marked in many places, formed contemporaneously with the minor folding. Fracture cleavage, which is essentially parallel to the axial planes of the minor folds, formed during the later stages of the deformation.

Metamorphism.—The metamorphism is syntectonic. Recrystallization began sometime after folding had commenced. The major structure must have already developed because the character of the metamorphism is

controlled by the larger structural features. Locally recrystallization was contemporaneous with a later phase of the orogeny, during which many minor folds developed. Elsewhere, however, crystallization, notably of andalusite, preceded this later phase of minor folding, but many of the other minerals crystallized afterwards. On the whole, recrystallization was syntectonic, but some preceded the intensive minor folding, some followed.

Some rocks recrystallized without any significant change in chemical composition, such as the amphibolite and biotite gneiss in the Ammonoosuc volcanics, the shiny pseudo-andalusite schist in the Partridge and Littleton formations, and probably the actinolite and diopside granulites of the Fitch formation—except for the loss of CO_2. Potash was added to some schists, notably the coarse andalusite schist, the coarse pseudo-andalusite schist, and the coarse rough pseudo-andalusite schist. Potash was added in two stages; in an earlier high temperature stage the aluminous minerals were converted in whole or in part to muscovite. In a later, lower temperature stage sericite replaced the unaltered andalusite, and chloritization of biotite and garnet took place. Magnesia was in part removed during the metamorphism.

The gneiss originated in part through selective solution *in situ*, accompanied by the introduction of soda, lime, and potash. Where the light-colored constituents constitute a high percentage of the gneiss, 4 or 5 per cent of soda, lime, and potash have been added to produce granitic-looking rocks.

The gneiss formed near the end of the orogeny, because some of the light-colored material has concentrated along cleavage planes parallel to the axial planes of the folds. The orogeny was in an advanced state when the segregation occurred; although the gneiss is locally thrown into small folds, the crystallization is later than the folding. The rocks show no evidence of strain, such as would be expected if they were folded subsequent to the crystallization of the light-colored material.

LATE-OROGENIC AND POSTOROGENIC HISTORY

After the main orogenic movements had ceased, the Bickford granite intruded the relatively brittle older rocks. It made space for itself partly by forcing aside the country rocks, but partly by stoping. Subsequently a series of northeasterly trending normal faults, the largest of which is the Pine Mountain fault developed. The last phase in the construction of the bedrock was the intrusion of the White Mountain magma series, to which may be assigned the syenite on Cherry Mountain, the Conway granite, tuffs and breccias in volcanic vents, and small dikes.

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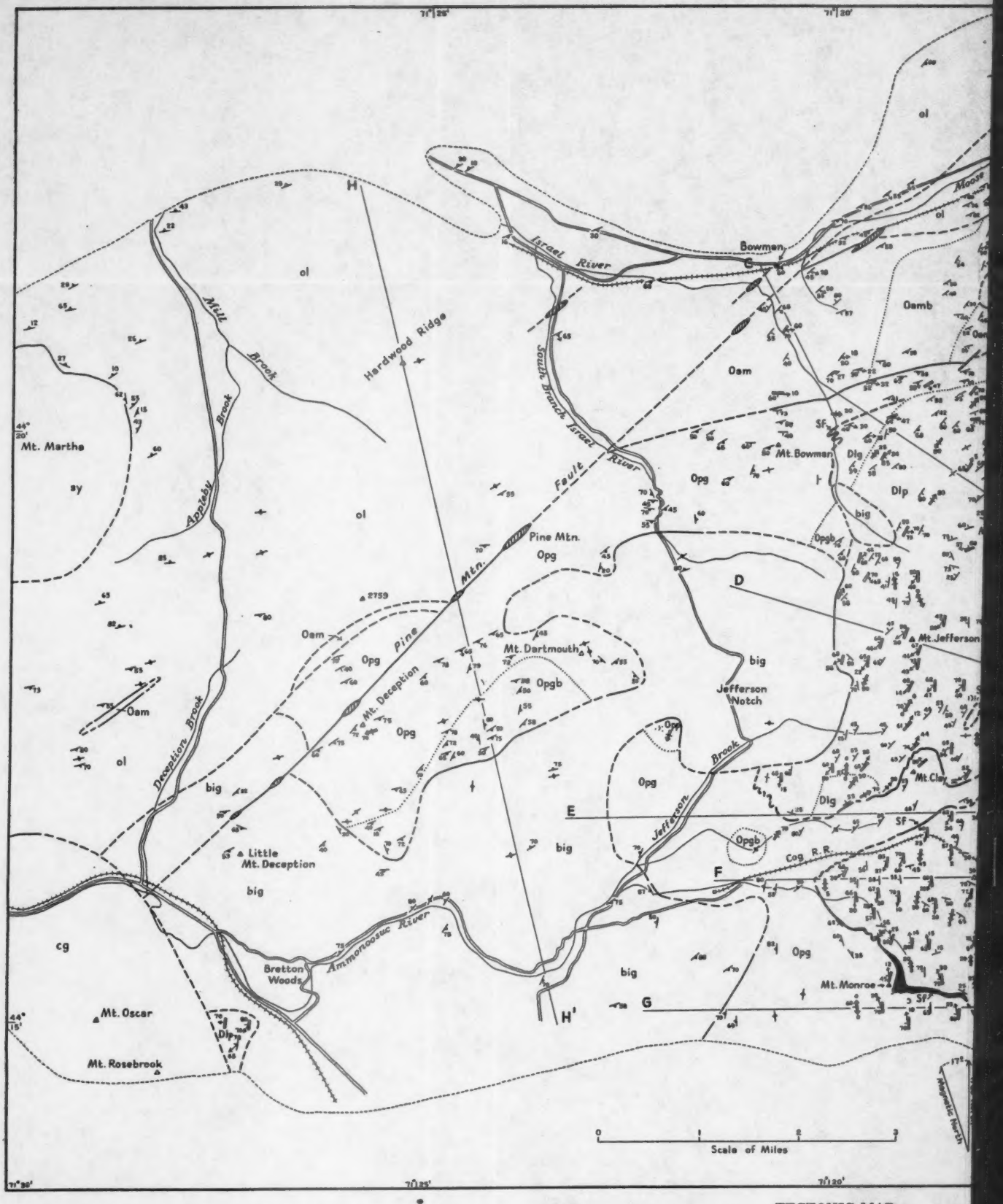
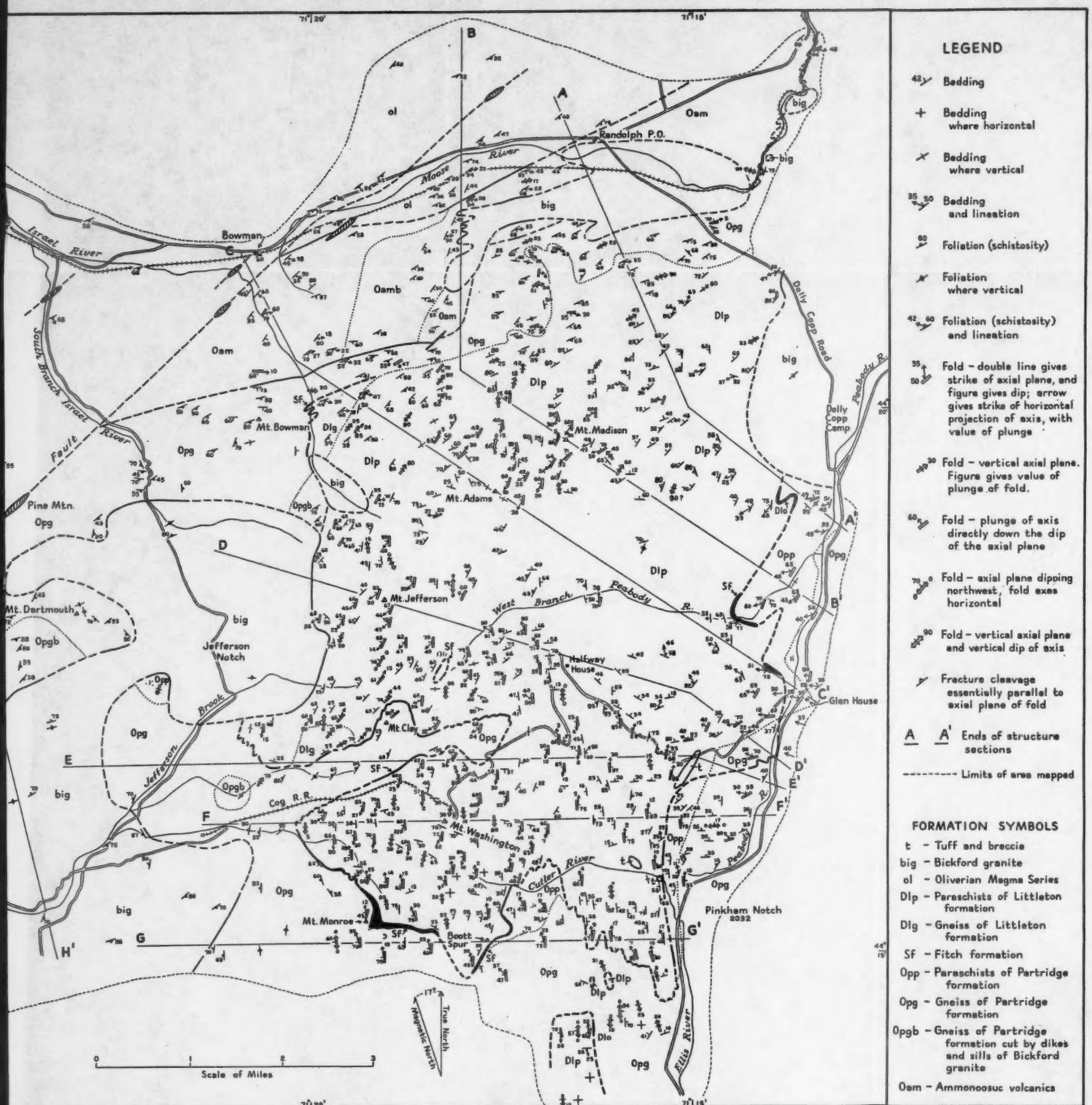


TECTONIC MAP

Shows by appropriate symbols structural data. Only part of the field observati



LEGEND

- Bedding
- Bedding where horizontal
- Bedding where vertical
- Bedding and lineation
- Foliation (schistosity)
- Foliation where vertical
- Foliation (schistosity) and lineation
- Fold – double line gives strike of axial plane, and figure gives dip; arrow gives strike of horizontal projection of axis, with value of plunge
- Fold – vertical axial plane. Figure gives value of plunge of fold.
- Fold – plunge of axis directly down the dip of the axial plane
- Fold – axial plane dipping northwest, fold axes horizontal
- Fold – vertical axial plane and vertical dip of axis
- Fracture cleavage essentially parallel to axial plane of fold
- A A' Ends of structure sections
- Limits of area mapped

FORMATION SYMBOLS

- t – Tuff and breccia
- big – Bickford granite
- ol – Oliverian Magma Series
- Dlp – Paraschists of Littleton formation
- Dlg – Gneiss of Littleton formation
- Sf – Fitch formation
- Opp – Paraschists of Partridge formation
- Opg – Gneiss of Partridge formation
- Opgb – Gneiss of Partridge formation cut by dikes and sills of Bickford granite
- Oam – Ammonoosuc volcanics

TECTONIC MAP

by appropriate symbols structural data. Only part of the field observations could be recorded on this scale.

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